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Late Cenozoic tectonics of the southern Inyo Mountains, Eastern California

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LATE CENOZOIC TECTONICS OF THE SOUTHERN INYO MOUNTAINS, EASTERN CALIFORNIA

A Thesis

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TABLE OF CONTENTS

LIST OF ILLUSTRATIONS

LIST OF TABLES

ABSTRACT

Two major range-bounding fault zones exhibiting large amounts of recent offset, the Eastern Inyo fault zone and the Hunter Mountain fault zone, intersect along the eastern flank of the southern Inyo Mountains in southeastern California. A previously unrecognized extension of the Eastern Inyo fault zone continues southward into Lee Flat. Along the eastern flank of the Inyo Mountains east of Cerro Gordo and north of Lee Flat, a section of older, dissected gravels, here termed the gravels of Bonham Canyon, overlie the continuation of the Eastern Inyo fault zone. Their position in this area suggests that they are alluvial fan deposits related to an early phase of uplift of the Inyo Mountains block along this fault zone.

The gravels of Bonham Canyon are about 365 m thick and contain abundant debris-flow deposits, characteristic of alluvial-fan deposits near their source. Paleocurrent directions in the gravels are generally eastward and define a mean transport direction of N. 52° E., roughly perpendicular to the trend of the Eastern Inyo fault zone. The gravels appear to be derived from an area in the Inyo Mountains underlain by a section of Ordovician to Mississippian sedimentary rocks characterized by Ordovician and Silurian dolomite and quartzite, Devonian limestone, and Mississippian siltstone and shale. Clast counts in the gravels show an abundance of siltstone and shale in the lower part of the section, a dominance of limestone in the middle part of the section, and increasing proportions of dolomite and quartzite in the upper part, suggesting the progressive downward erosion of the uplifted Inyo Mountains block.

Biotite from a water-worked tephra bed about 5 m from the bottom of the gravels was dated by K-Ar and $^{40}Ar/^{39}Ar$ techniques, giving an age of 13.6±0.5 Ma. This age marks the beginning of deposition of the Bonham Canyon gravels and suggests that uplift

of the Inyo Mountains began at about 14 Ma, about the same time as the initiation of major extension in Death Valley. The northwest-trend of the Eastern Inyo fault zone suggests ENE-WSW-oriented extension. This episode of faulting ceased in the Inyo Mountains region by late Miocene time when a mature erosion surface developed.

After about 4 Ma, renewed extension, this time oriented in an ESE-WNW direction, resulted in resumed normal faulting along the northern part of the Eastern Inyo fault zone and oblique slip on the Hunter Mountain fault zone. Modern Saline Valley formed at this time. South of Saline Valley, the gravels of Bonham Canyon remained stranded astride the southern part of the ancestral Eastern Inyo fault zone, whereas to the north coeval deposits were either down-dropped into Saline Valley or were eroded.

An area of several square kilometers south of the Hunter Mountain fault zone is underlain by shattered and broken upper Paleozoic limestone and shale. This tectonic breccia, termed the "jumble," is attributed to fracturing and sliding of these sedimentary rocks toward Saline Valley. The brittle nature of the shattering suggests that deformation occurred at or near the surface, implying that the deformation is relatively recent and is genetically related to the range-front faults. This deformation dies out southwards, where the Paleozoic rocks are overlain by the gravels of Bonham Canyon.

The late Cenozoic tectonic history of the southern Inyo Mountains in conjunction with data from surrounding regions suggests that extension in the Inyo-Death Valley region occurred in two distinct periods with different extension directions. The superposition of post-4 Ma extension on middle Miocene structures that formed in response to a different extension direction is responsible for the modern topography of the Inyo-Death Valley region.

INTRODUCTION

Investigation of late Cenozoic features of the southern Inyo Mountains was undertaken in order to better delineate the history and development of extensional basin and range faulting in the Inyo-Death Valley region (Fig. 1). This region comprises a complex network of northeast- and northwest-striking faults with variable amounts of offset and sense of slip, which offer a striking contrast to other parts of the Great Basin, which typically show a relatively simple physiography of north-trending mountain blocks and intervening basins. Studies in the southern Inyo Mountains, when integrated with other studies of the Inyo-Death Valley region, provide new insight into the general late Cenozoic tectonic history of the region.

The area of this study was selected because it contains a number of important features related to late Cenozoic tectonics. These features have received little attention by previous workers, largely because of difficult access and the rugged topography of the southeastern Inyo Mountains. Two well-recognized structures that nonetheless have been little studied in the context of the late Cenozoic tectonic development of the southern Inyo Mountains are the Eastern Inyo and Hunter Mountain fault zones (Fig. 2). Another feature of importance is a unique and previously unrecognized tectonic breccia, termed the "jumble" by Conrad and McKee (1985), that underlies a graben-like depression immediately south of the intersection of the Eastern Inyo and Hunter Mountain fault zones. Also present is a wedge-shaped deposit of older, dissected conglomerate, here termed the gravels of Bonham Canyon, that appears to represent an important phase of basin development in the southern Inyo Mountains.

This investigation of the late Cenozoic tectonic development of the southern Inyo Mountains included detailed mapping within the area of study to delineate important structures and relationships and to establish links between the development of late

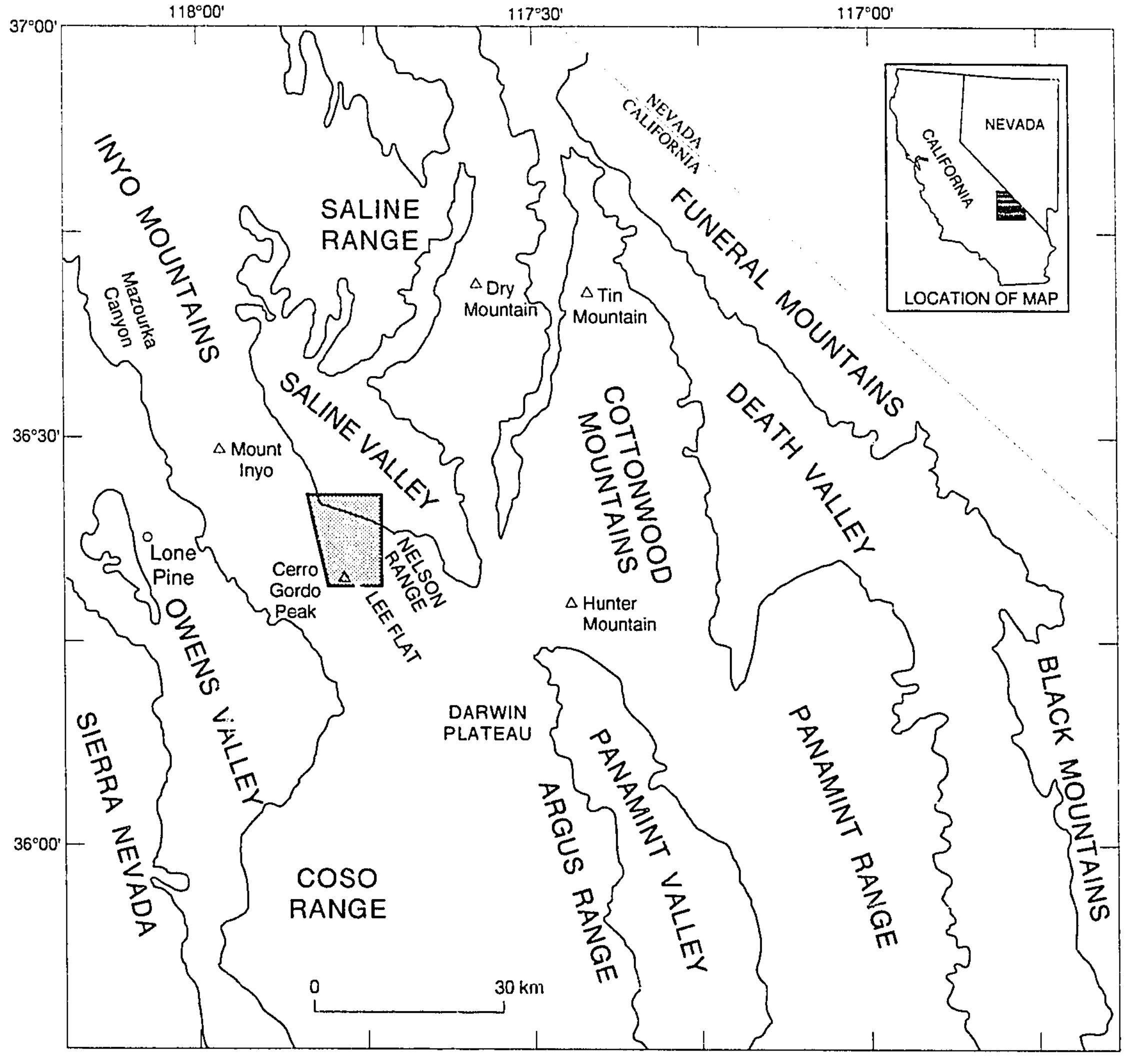

Figure 1. Index map showing area of study (shaded) and principal physiographic features in the Inyo-Death Valley region.

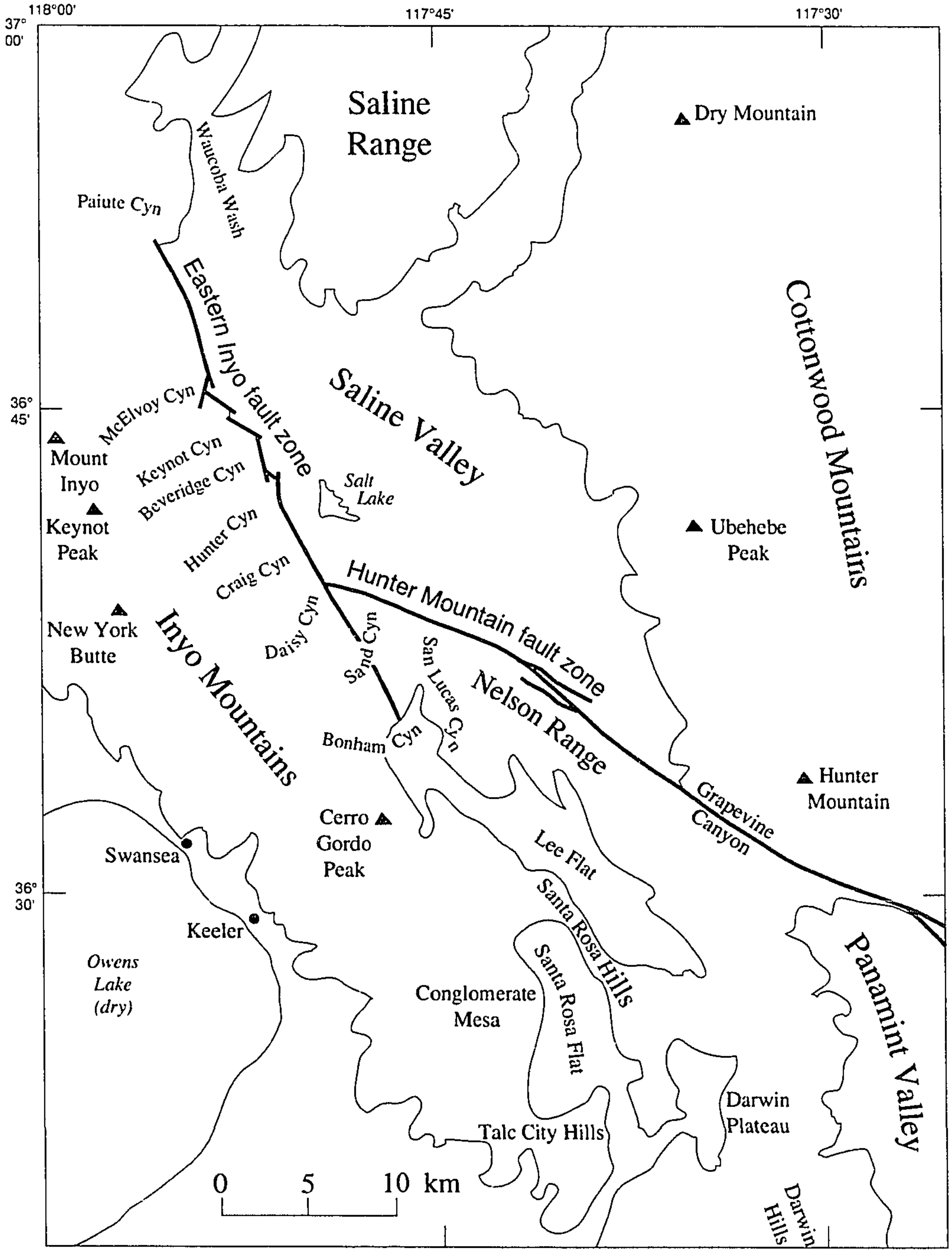

Figure 2. Principal physiographic features of the southern Inyo Mountains.

Cenozoic structures, formation of the jumble, and deposition of the gravels of Bonham Canyon. Reconnaissance mapping in adjacent parts of the Inyo Mountains served to better characterize the age of some features and helped to distinguish pre-Cenozoic from late Cenozoic deformation. Sedimentologic and geochronologic studies of the gravels of Bonham Canyon were used to define the source, significance, and age of this deposit. All of the data collected was used to interpret the late Cenozoic structural history of the southern Inyo Mountains, which has important implications for the understanding of late Cenozoic tectonics for the entire Inyo-Death Valley region.

Location and Physiography

The part of the Inyo Mountains included in the study area lies on the eastern side of the range east of Cerro Gordo Peak, along the southwestern margin of Saline Valley (Fig. 2). The terrain is extremely rugged, total relief being greater than anywhere else in the Great Basin; in less than 11 km, elevation rises from 350 m at the floor of Saline Valley to 3385 m at the top of Mount Inyo. The climate is arid to semiarid. Vegetation consists of scattered sage brush, greasewood, rabbit brush, and grasses at lower elevations and piñon pine, juniper, and mountain mahogany at higher elevations.

Access to much of the southern Inyo Mountains is limited to trails. Dirt roads, including a well-maintained road that connects the town of Keeler to the Cerro Gordo mine near the ridge crest and extends down the east side of the range to the upper part of San Lucas and Bonham Canyons, provide access to parts of the southern part of the range. An unimproved dirt road that extends from the Cerro Gordo mine to the Burgess mine along the Inyo crest and down into the upper part of the Hunter Canyon provides access along the upper part of the Inyo Mountains. A jeep trail from Swansea in Owens Valley also connects to this road. From Saline Valley, the well-maintained Saline Valley road provides access to range front areas on the southern and western margins of Saline Valley and unimproved dirt roads branch off of it to the mouths of several canyons.

Previous Studies

Published geologic maps of parts of the southern Inyo Mountains at a scale of 1:62,500 include the Independence quadrangle (Ross, 1965), the Waucoba Wash quadrangle (Ross, 1967), and the Ubehebe Peak quadrangle (McAllister, 1956). Geologic maps of parts of the New York Butte quadrangle, based in part on unpublished mapping by W.C. Smith of the U.S. Geological Survey, were published by Conrad and McKee (1985) and Conrad and others (1987). Pre-Cenozoic rocks in part of the southern Inyo Mountains and areas to the east have been mapped at a scale of 1:32,500 by Stone and others (1989), and part of the area of study was mapped at a scale of 1:12,000 in a study of the stratigraphy and structure of the southeastern Inyo Mountains by Husk (1979).

Early work in the Inyo Mountains includes that of Knopf (1914; 1918) and Kirk (*in* Knopf, 1918), who first described the stratigraphy and published reconnaissance geologic maps of parts of the range. The Paleozoic and Triassic sedimentary rocks in the area of study were first described in detail by Merriam (1963) in studies related to the Cerro Gordo mine. The stratigraphy of Cambrian, Ordovician, and Silurian rocks was subsequently described by Ross (1963; 1965; 1966). Mississippian rocks have been described by Stevens and others (1979), Dunne and others (1981), and Klingman (1987). Studies of Pennsylvanian and Permian rocks include those by Merriam and Hall (1957), Stone (1984), Stone and Stevens (1984; 1987; 1988a), and Stone and others (1989). Triassic rocks have been studied by Smith (1914; 1932), Dunne and others (1978), Oborne (1983), Stone and Stevens (1986), and Stone and others (1991). Stevens (1986; 1991) summarized of the Ordovician to Triassic geologic history of eastern California.

Descriptions of the Mesozoic calc-alkalic and alkalic plutonic rocks in the southern Inyo Mountains have been given by Bateman and others (1963), McKee and Nash (1967), Ross (1969), Miller (1977; 1978), Sylvester and others (1978), Dunne (1979; 1986), Griffis (1986), and Dunne and Walker (1991). Moore and Hopson (1961), Chen and Moore (1979), and Dunne (1986) have studied and interpreted the Independence dike swarm.

The structural and tectonic evolution of the region has been studied by many workers. Evidence for Paleozoic deformation has been described by Sylvester and Babcock (1975), Stevens and Stone (1988), Stone and Stevens (1988b), and Stone and others (1989). Mesozoic contractile deformation, characterized by major thrust faulting and folding, is widespread in the region. The Last Chance thrust and related thrust faults have been described by Stewart and others (1966), Corbett and others (1988), and Snow (1992). Other related compressional features include the Inyo thrust fault (Stevens and Olson, 1972), and the East Sierran thrust system, which has been studied by Kelley and Stevens (1975), Gulliver (1976), Dunne and Gulliver (1976), and Dunne and others (1978; 1983). A summary of Mesozoic deformation in the region was given by Dunne (1986).

Studies of Cenozoic structures and tectonics in and near the area of study include an examination of the recent deformation in Saline Valley by Zellmer (1980) and the evolution of the Hunter Mountain fault zone by Burchfiel and others (1987) and Sternlof (1988). Schweig (1982; 1989) described the late Cenozoic tectonics and volcanic rocks of the Darwin Plateau. Other studies of extension-related volcanic rocks in the Darwin Plateau and Saline Range include those by Ross (1970), Larsen (1979), and Coleman and Walker (1990). Numerous studies of late Cenozoic tectonics in the region have centered in Death Valley and the Panamint Range but have important regional implications. These include works by Stewart (1983), Cemen and others (1985), Wernicke and others (1988), Snow and White (1990), McKenna and Hodges (1990), and Hodges and McKenna (1990).

LITHOLOGIC UNITS

The area of study is underlain by a sequence of intensely folded and faulted marine sedimentary rocks of Ordovician through Permian age with an approximate total thickness of nearly 3,000 m (Table 1). Ordovician and Silurian rocks are mostly dolomitic with minor quartzite, Devonian rocks are comprised mainly of dolomite and limestone, Mississippian rocks are dominated by shale and argillite with minor limestone, and Pennsylvanian and Permian rocks are composed of silty limestone. These rocks have been intruded by granitic rocks of Mesozoic age. Late Cenozoic volcanic and sedimentary rocks, including at least 900 m of sedimentary rocks in Saline Valley and lesser amounts in the vicinity of Lee Flat (Fig. 2), unconformably overlie the older rocks. Most of the pre-Cenozoic rocks have been described in detail previously (Merriam, 1963; Ross, 1965; Husk, 1979; Dunne and others, 1978; Dunne, 1979; Stevens, 1991); brief summaries are presented here.

Paleozoic Sedimentary Rocks

Badger Flat Formation

The Badger Flat Formation includes the oldest rocks exposed in the study area. The base of the formation is not exposed, but 150-175 m of the section is present in Bonham Canyon (Plate 1). The Badger Flat Formation, recognized throughout the Inyo Mountains and the Talc City Hills (Fig. 2), was named for exposures near Mazourka Canyon (Fig. 1) in the western Inyo Mountains (Ross, 1963). Rocks of the Badger Flat Formation are a lateral equivalent of the Antelope Valley Formation of the Pogonip Group, a unit recognized in areas east of the Inyo Mountains and named for exposures near Eureka, Nevada (Nolan and others, 1956).

Table 1. Stratigraphic column for the southeastern Inyo Mountains

Age	Formation	Thickness (m)	Lithology
Quaternary	Surficial Deposits	200+?	Unconsolidated alluvium, fanglomerate, landslides, and talus deposits
	—Unconformity—		
Pliocene	Basalt	200	Olivine basalt, basaltic lapilli tuff and tuff breccia, and alluvial deposits
	—Unconformity—		
Miocene	Gravels of Bonham Canyon	365	Pebble, cobble, and boulder conglomerate
	—Unconformity—		
Permian Pennsylvanian	Keeler Canyon Formation	760	Sandy limestone, shale, and siltstone
Mississippian	Rest Spring Shale	400	Dark-gray, silty shale
	Perdido Formation	75	Limestone, shale, and quartzite
	Tin Mountain Limestone	105	Dark blue-gray, cherty limestone
Devonian	Lost Burro Formation	488	Light- to dark-blue-gray limestone
Devonian Silurian	Hidden Valley Dolomite	535	Massive, tan to light-gray dolomite, cherty dolomite, and quartzite
Ordovician	Ely Springs Dolomite	75	Light- and dark-gray, cherty dolomite
	Eureka Quartzite	60	Light-gray, vitreous quartzite
	Badger Flat Formation	175+	Medium- to light-gray dolomite
	—Base not exposed—		

In contrast to the type area, where the formation consists mostly of dark- to medium-gray limestone, rocks of the Badger Flat Formation in the southern Inyo Mountains are comprised of iron-stained, medium- to dark-blue-gray, mottled dolomite. Despite alteration and recrystallization, several types of fossils are preserved including the alga *Receptaculites* and the gastropods *Maclurites* and *Palliseria*. These fossils typify the Badger Flat Formation and suggest an Early to Middle Ordovician age (Merriam, 1963).

Eureka Quartzite

The Eureka Quartzite conformably overlies the Badger Flat Formation and is exposed in Bonham and San Lucas Canyons (Plate 1). The unit was named for exposures in the Eureka district in central Nevada (Nolan and others, 1956) and is exposed throughout the western Great Basin. The Eureka Quartzite is the lateral equivalent of limestone, dolomite, and quartzite of the Johnson Spring Formation, which is exposed in the Mazourka Canyon area (Fig. 1) and marks the southern Inyo Mountains as an area transitional between the more offshore Mazourka facies to the northwest and nearshore facies to the east (Stevens, 1986). The unit is about 60 m thick in the study area.

The Eureka Quartzite is composed of clean, dense, vitreous, thick-bedded to massive quartzite. The unit is typically nearly white in color, although weathered outcrops locally are stained various shades of gray, brown, black, or red. Where the unit is not highly recrystallized, crossbeds up to 1 m thick are common. The Eureka Quartzite is considered to be Middle Ordovician in age based on stratigraphic position and sparse fossil evidence (Ross, 1965).

Ely Springs Dolomite

The Ely Springs Dolomite is exposed in San Lucas Canyon where it conformably overlies the Eureka Quartzite (Plate 1). The formation was named for exposures in southeastern Nevada (Westgate and Knopf, 1932) and is widely exposed throughout the southern and central Great Basin (Merriam, 1963). The formation is at least 75 m thick in the study area, although the top of the formation is not easily defined because the overlying Silurian rocks are of similar lithology. Merriam (1963) defined this upper transitional contact on a paleontologic basis. Although fossils are scarce in the study area due to alteration, those from the Talc City Hills (Fig. 2) about 25 km to the southeast indicate a Late Ordovician to Early Silurian age for the formation (Miller, 1975).

The Ely Springs Dolomite is composed of dark-blue-gray, thin-bedded, cherty saccharoidal dolomite. Abundant chert occurring as nodules and lenses is characteristic of this unit and makes up as much as 30% of the rock. The upper part of the formation contains much less chert and is mainly blue-gray, medium- to thick-bedded, coarse-grained dolomite that passes gradationally into tan and light-gray rocks mapped as the Hidden Valley Dolomite.

Hidden Valley Dolomite

The Hidden Valley Dolomite crops out in the area of Bonham and San Lucas Canyons (Plate 1). Nearly half the unit consists of tan, light-gray, or white, thick-bedded, blocky dolomite. The remainder is composed of cherty, zebra-striped dolomite, and quartzite. The dolomite is locally altered to talc. Merriam (1963) reported a thickness of 535 m (1,750 ft) in Bonham Canyon.

Ross (1967) considered the Hidden Valley Dolomite to be Silurian and Devonian in age on the basis of sparse coral, algal, and brachiopod faunas.

Lost Burro Formation

Conformably overlying the Hidden Valley Dolomite is the Lost Burro Formation, exposed in the southern Inyo Mountains from near Cerro Gordo Peak to Daisy Canyon (Plate 1). The unit generally follows the axial trace of the Cerro Gordo anticline; within the study area, the unit is strongly foliated by bedding-parallel shear related to this fold. In Craig Canyon (Fig. 2), the formation forms spectacular 600-m cliffs that exhibit contorted blue and white banding. Merriam (1963) measured a thickness of 488 m (1,600 ft) for the unit at Cerro Gordo Peak.

The Lost Burro Formation is composed of thick-bedded white and blue-gray banded limestone with thin- to thick-bedded quartzite layers near its base. In contrast to the type section near Ubehebe Peak (Fig. 2), where the formation is mostly dolomite, and to exposures in the Talc City Hills (Fig. 2), which consist of roughly equal amounts of dolomite and limestone, dolomite occurs only at the base of the section in the vicinity of Cerro Gordo Peak (Merriam, 1963). A sandy dolomite containing abundant chert in the lower part at the type section near Ubehebe Peak (McAllister, 1952) and a 20-m-thick quartzite that marks the base of the formation in the Talc City Hills (Stone and others, 1989) are not recognized in the vicinity of Cerro Gordo. In this area, the base of the unit is placed at the bottom of a zone of partially dolomitized, blue-gray limestone (Merriam, 1963). The Lost Burro Formation is considered to be Middle to Late Devonian on the basis of abundant coral, stromatoporoid, and brachiopod faunas (Nelson, 1971).

Tin Mountain Limestone

The Tin Mountain Limestone crops out at the head of Sand Canyon and in a discontinuous band extending from the range crest at Cerro Gordo (Plate 1) to the north side of Craig Canyon, where it has been folded into small inliers within the Lost Burro Formation (Conrad and McKee, 1985). At Cerro Gordo Peak the Tin Mountain

Limestone is 105 m (350 ft) thick (Merriam, 1963), but north of here the unit is less than about 25 m (75 ft) thick and appears to pinch out in the vicinity of Craig and Hunter Canyons. The unit also is missing in the upper part of Sand Canyon because of erosion, nondeposition, or faulting.

The Tin Mountain Limestone is a medium- to dark-blue-gray, fine-grained limestone containing many dark-gray or black chert nodules and lenses. The uniform dark-blue color and abundant chert readily distinguishes this formation from the much lighter, banded Lost Burro Formation. Beds in the formation range from less than 1 cm to 0.6 m in thickness. The unit contains abundant fossils, chiefly crinoid and coralline debris.

The Tin Mountain Limestone was assigned an Early Mississippian age by Merriam (1963). It lies with apparent conformity on the Lost Burro Formation, but it is missing the shaly beds at its base that were noted by McAllister (1952) at the type section. Northwest of Cerro Gordo, the Tin Mountain is unconformably overlain by the Mississippian Perdido Formation.

Perdido Formation

The Perdido Formation consists of a heterogeneous sequence of strata that includes siltstone, shale, quartzite, and limestone. It is exposed south of San Lucas Canyon and in a small isolated outcrop about 3 km north of Bonham Canyon (Plate 1), where it is comprised of well-bedded, gray quartzite with interbeds of tan to light-gray limestone. In the vicinity of Cerro Gordo Peak and to the north, the unit commonly ranges from about 15 to 25 m (50-75 ft) in thickness, but in many places it is missing (Merriam, 1963). The formation occurs as two isolated outcrops of dark quartzite at the base of the Rest Spring Shale in Hunter Canyon, and south of San Lucas Canyon it consists of 50 to 75 m of light-brown, platy siltstone with thin interbeds of medium-gray limestone (Stone and others, 1989). Fossils at the type locality indicate a Late Mississippian age for the formation (McAllister, 1952).

Rest Spring Shale

In the southern Inyo Mountains, the Rest Spring Shale is exposed in two discontinuous bands, one along each limb of the Cerro Gordo anticline (Plate 1). One band extends from the top of Sand Canyon and Bonham Canyon northward to Craig Canyon in the rugged lower slopes of the range. The second band extends north from the range crest at Cerro Gordo to McElvoy Canyon. The thickness of the Rest Spring Shale varies considerably in the study area due to faulting and folding, ranging in thickness from 15 m or less north of Daisy Canyon to as much as 400 m in Craig Canyon.

The Rest Spring Shale is composed of dark-gray to black claystone, with minor interbedded silty shale, fine sandstone, and limestone. The claystone ranges from fissile shale to a nonfissile, dense, platy to blocky argillite. In the southern part of the Inyo Mountains, these shales locally have been recrystallized to slate, phyllite, fine-grained hornfels, and andalusite-bearing hornfels. Husk (1979) mapped a section of sheared, isoclinally folded siltstone and shale in Sand Canyon (Plate 1) as the Hamilton Canyon Formation of Langenheim (1962), but because of the deformation no attempt was made to separate this unit from the Rest Spring Shale as used in this report. Scattered fossils and stratigraphic position suggest that the Rest Spring Shale is Late Mississippian in age (Merriam, 1963).

Keeler Canyon Formation

The Keeler Canyon Formation conformably overlies the Rest Spring Shale. It is exposed along the crest of the Inyo Mountains from Cerro Gordo Peak to Keynot Peak (Fig. 2) and along the lower part of the range from Beveridge Canyon to the upper part of

Sand Canyon (Fig. 2; Plate 1). It is also exposed in Sand Canyon (Plate 1) and to the east in the Nelson Range (Fig. 2). In the Cerro Gordo area, Merriam (1963) reported the formation to be 400 m to 760 m (1,300 ft to 2,500 ft) thick, but in most places the thickness is difficult to determine because of folding and faulting.

The Keeler Canyon Formation is a turbidite sequence composed of medium- to thick-bedded, dark-gray, silty and arenaceous to pebbly, graded, bioclastic limestone intercalated with gray to pink shale. The lower 50 to 75 m of the formation contains aphanitic, medium-gray limestone with spheroidal, black and brown, chert nodules 1 to 4 cm in diameter. Known informally as the "golf ball beds," this part of the formation provides a stratigraphic marker also recognized in the Darwin Hills, the Argus Range, and the Panamint Range 50 km to the east (Merriam, 1963). Based on abundant fusulinids in the Argus and Panamint Ranges, the formation is dated Middle Pennsylvanian to Early Permian in age (Hall, 1971).

Mesozoic Plutonic Rocks

Mesozoic plutonic rocks underlie a large part of the central Inyo Mountains and many smaller stocks are present in the southern part of the range. Three different suites of granitic rocks are recognized in the region (Dunne, 1986). These are: 1) a calc-alkalic suite consisting mostly of granite and granodiorite that is considered to be satellitic to the Sierra Nevada batholith (Bateman and others, 1963; McKee and Nash, 1967; Ross, 1969; Dunne, 1986); 2) an alkalic suite that forms a discontinuous belt along the east side of the Sierra Nevada batholith (Miller, 1978) and is considered coeval, but not comagmatic, with rocks of the Sierra Nevada batholith (Griffis, 1986); and 3) a peraluminous suite consisting of small, leucocratic, high-silica plutons composed of granite and monzogranite (Dunne, 1986; Griffis, 1986; Griffis and others, 1986). All of these suites are present in the southern Inyo Mountains, but within the area of this study, rocks of the alkalic suite, representing outliers of the Hunter Mountain batholith, are the main type of intrusive rock. Dioritic rocks that may belong to the calc-alkalic suite are of secondary importance, and rocks of the peraluminous suite have not been recognized in the study area. Also present in the southern Inyo Mountains are numerous, mostly northwest-trending, steeply dipping dikes that are considered part of the Independence dike swarm (Moore and Hopson, 1961; Chen and Moore, 1979; Dunne, 1986).

Hunter Mountain Quartz Monzonite

Rocks considered part of the Hunter Mountain batholith, a large representative of the alkalic intrusive suite, underlie a large part of the central Inyo Mountains (Ross, 1965; 1969; Conrad and McKee, 1985). Smaller stocks related to it are present at New York Butte (Fig. 2) and west of Sand Canyon (Plate 1). The composition ranges from quartz monzonite to diorite, but quartz monzonite and granodiorite are predominant. The central core of the pluton in the Inyo Mountains consists of a medium-gray, medium-grained, seriate quartz monzonite with pinkish to pale-red K-feldspar phenocrysts ranging up to 1 cm in length. These rocks grade into a dark, fine- to medium-grained diorite (Ross, 1969) in the eastern and southern parts of the pluton. The quartz monzonite is made up of approximately 35 percent plagioclase, 30 percent K-feldspar, 20 percent quartz, and 15 percent hornblende and subordinate biotite; the dark diorite consists of about 50 percent plagioclase, 17 percent K-feldspar, 7 percent quartz, and 25 percent hornblende with minor amounts of magnetite, apatite, sphene, and zircon. Radiometric ages from the central part of the Hunter Mountain batholith include U/Pb ages of 174 to 180 Ma on zircon and a biotite K-Ar age of 185 Ma, indicating that the Hunter Mountain Quartz Monzonite is Early Jurassic in age (Sylvester and others, 1978; Chen and Moore, 1982; Dunne, 1986).

Dioritic Rocks

Diorite of inferred Jurassic age is exposed along the eastern front of the Inyo Mountains from Keynot Canyon to Hunter Canyon (Fig. 2; Conrad and McKee, 1985). Several smaller diorite stocks include those exposed in Daisy Canyon, Sand Canyon, and San Lucas Canyon (Plate 1). The diorite typically is medium to dark gray, medium grained, and composed of 40-60 percent plagioclase, 5-20 percent K-feldspar, 0-20 percent hornblende, 0-15 percent biotite, and 0-10 percent quartz. These bodies may be part of either the calc-alkalic suite or the alkalic suite.

Independence Dike Swarm

A number of light- to dark-colored dikes which are probably part of the Independence dike swarm (Moore and Hopson, 1961) occur throughout the study area (Plate 1). The dikes, most of which are nearly vertical, generally trend north to northwest. The light-colored dikes and stocks have a variety of textures and compositions including alaskite, pegmatite, aplite, granite, quartz monzonite, and granodiorite. In Bonham Canyon, several greenish-gray, porphyritic, andesitic to dacitic dikes crop out that have been assigned to this assemblage (Merriam, 1963). The Independence dike swarm is considered to be Late Jurassic in age, based on a U/Pb zircon age of 148 Ma (Chen and Moore, 1979).

Cenozoic Volcanic and Sedimentary Rocks

The Gravels of Bonham Canyon

The gravels of Bonham Canyon form a more or less crescent-shaped outcrop pattern east of Cerro Gordo Peak and north of Lee Flat (Fig. 2; Plate 1). These sediments were referred to by Merriam (1963) as the San Lucas fan. They consist of moderately- to well-cemented, boulder, cobble, and pebble conglomerate that weathers to smooth, gravel-covered slopes (Fig. 3). Outcrops are rare. These gravels have been considered to be older Quaternary gravels in other studies (McAllister, 1956; Husk, 1979; Conrad and McKee, 1985), but the deep dissection of these deposits suggests they may be considerably older.

Basalt

Olivine basalt flows and associated pyroclastic and sedimentary deposits are exposed several kilometers southeast of the study area in the vicinity of Lee Flat, Santa Rosa Flat, and the Darwin Plateau (Fig. 2). The composite thickness of these deposits exceeds 200 m on the eastern edge of the Darwin Plateau (Schweig, 1982). The basalt commonly contains about 20 percent phenocrysts composed of subequal amounts of olivine and plagioclase in an aphanitic groundmass of plagioclase and olivine (Schweig, 1982). Some flows contains xenocrysts of partially resorbed quartz (Hall, 1971; Schweig, 1982). K-Ar ages range from about 7.7 to 4.3 Ma for rocks in the Darwin Plateau (Hall, 1971; Larsen, 1979; Schweig, 1982). Chemically similar basalt is exposed north of Saline Valley in the Saline Range. K-Ar ages for these rocks are younger, ranging from 4.6 to 1.7 Ma (Ross, 1970; Larsen, 1979; Elliott and others, 1984).

Quaternary Surficial Deposits

Quaternary sediments in the study area are divided into older and younger deposits. The older deposits typically form small, discontinuous outcrops in canyon bottoms or perched remnants on the sides of valleys. Small deposits also occur on the uplifted sides of high-angle, basin-and-range faults along the range front. These units generally consist of well bedded and partially consolidated sand and gravel and include fanglomerate and talus.

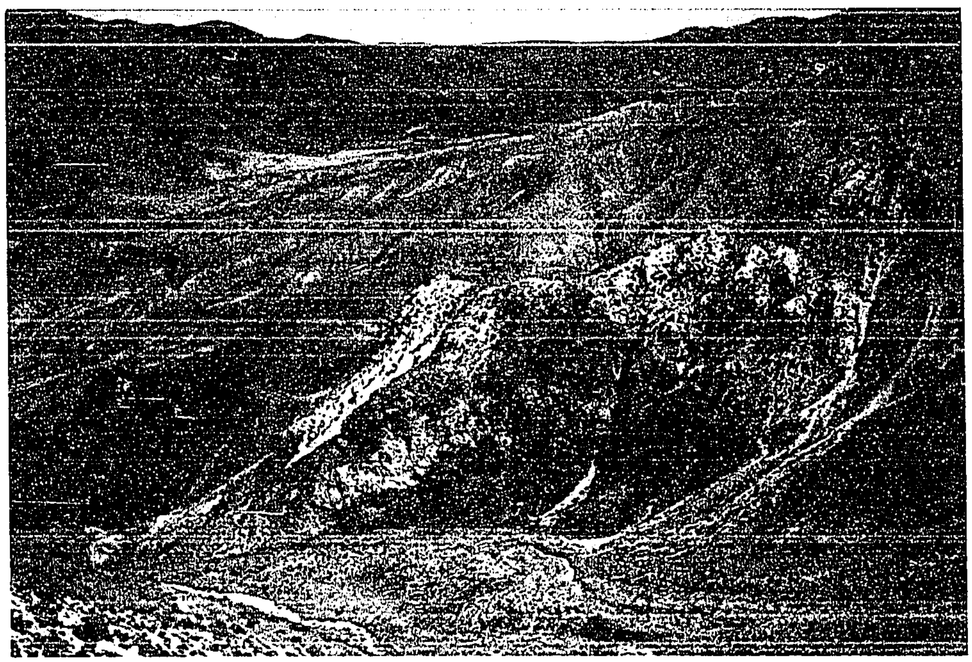

Figure 3. View southward of the gravels of Bonham Canyon. The gravels underlie the smooth, gray and tan slopes in the middle distance. The banded blue and white ridge in foreground, underlain by the limestone of the Lost Burro and Perdido Formations, is about 100 m high. Note the shallow eastward dip of the gravels in the outcrops beyond the limestone ridge.

The younger Quaternary deposits consist of unconsolidated fanglomerate, alluvium, and colluvium. The fanglomerate occurs primarily along the edge of Saline Valley, the alluvium occurs along stream bottoms and in the bottom of Saline Valley, and the colluvium consists of rock debris and talus on the slopes.

Landslides

There are several large and many small landslides along the steep eastern slopes of the Inyo Mountains and some steep slopes show evidence of pull-away structures indicating the probability of future slides (Conrad and McKee, 1985). These large gravity features probably are due to recent rapid uplift of the Inyo Mountains along basin-and-range faults.

The most notable landslide along the range front is a 4 km^2, lobe-shaped mass of rock debris west of Salt Lake (Figs. 2, 4). Many smaller landslides occur in steep canyons and elsewhere on the eastern side of the range. A pull-away structure north of Hunter Canyon consists of a series of wedge-shaped blocks, detached from the bedrock, that have slipped down to form a series of about 12 terraces on the mountain face (Fig. 4). This structure is about 2,000 m long and 400 m wide and extends almost 1,000 m vertically.

A landslide located about 1 km north of Bonham Canyon on the eastern side of the map area (Plate 1) consists entirely of granitic rubble. It lies on rocks of the Keeler Canyon Formation and has no obvious local source. The nearest exposed granitic rocks are about 2 km to the west along the steep flank of the Inyo Mountains where at least two other landslides have been identified. This deposit may be a distal toe of one of these landslide masses that has since been isolated by erosion.

Figure 4. Landslide and pull-away structures along the eastern range front of the Inyo Mountains west of Salt Lake playa (Fig. 2). A series of about 12 terraces composed of wedge-shaped blocks that are detaching from the mountain face are visible in the center of the photograph. Dark-brown hill on the lower left is a landslide mass about 2 km wide.

STRUCTURE

Paleozoic and Triassic strata in the southern Inyo Mountains have been strongly folded and faulted during several episodes of contractional deformation ranging in age from Permian to possibly early Cenozoic. The intrusion of large plutons has further complicated the structure by causing widespread recrystallization and locally obscuring older structural relationships along major faults and other zones of crustal weakness. Late Cenozoic deformation is characterized by extensional basin-and-range faulting. Because this paper focuses on late Cenozoic structure and tectonics and because pre-Cenozoic structures in the region have been studied in detail by others (Dunne and others, 1978; Dunne, 1986; Stone and others, 1989), only a brief summary of pre-Cenozoic structures is presented here.

Pre-Cenozoic Structures

Permian Structures

Structures associated with Permian deformation have been identified by Stevens and Stone (1988). The oldest of these is the Fishhook thrust, a folded thrust fault exposed east of Conglomerate Mesa (Fig. 2). This fault and the related Lee Flat thrust, which is nowhere exposed but is inferred to exist under Cenozoic rocks between Conglomerate Mesa and the Santa Rosa Hills (Fig. 2), involve strata of Mississippian to earliest Permian age. Deformation which involved folding and several kilometers of displacement on these faults apparently terminated in Early Permian (mid-Wolfcampian) time (Stevens and Stone, 1988).

The development of broad, north- to northeast-trending folds during late Early Permian (Leonardian) or early Late Permian (Guadalupian) time is suggested by a regional unconformity developed on rocks of the Permian Owens Valley Group (Stone, 1984; Stone and Stevens, 1984). In places, up to 1,000 m of strata beneath the unconformity was eroded during or subsequent to deformation (Stone and others, 1989).

Mesozoic Structures

The Inyo-Death Valley region lies within a belt of contractional deformation characterized by the development of numerous major thrust faults and folds. The evolution of this belt is very complex and appears to be related to an oblique intersection of the Cordilleran foreland fold and thrust belt and the Sierra Nevada batholith (Dunne, 1986). Deformation probably began in Middle to Late Triassic time (Oborne and Dunne, 1982). Dunne (1986) divided Mesozoic structures in the region into three assemblages. These are: 1) the Last Chance assemblage; 2) the East Sierran assemblage; and 3) the Laramide assemblage.

Structures of the Last Chance assemblage include the Last Chance thrust fault (Stewart and others, 1966) and related thrust faults and associated folds. The thrust faults in this group commonly have large displacements estimated to range from 7 to 35 km. Transport directions are between N. 50° E. and S. 70° E. These faults comprise the southwesternmost portion of the Cordilleran thrust belt. Deformation associated with this assemblage probably began in the Middle to Late Triassic (Oborne and Dunne, 1982) and terminated by Middle Jurassic time (Dunne, 1986), a view also held by Stone and Stevens (1993). A contrasting interpretation is offered by Snow (1992) and Snow and Wernicke (1993), who consider the Permian Lee Flat and Fishhook thrusts of Stone and others (1989) to be part of the Last Chance thrust system, thereby suggesting that major crustal shortening began in Permian time.

The East Sierran assemblage consists of the East Sierran thrust system (Dunne and others, 1983) and associated folds that lie along the eastern margin of the Sierra Nevada batholith east of Owens Valley (Fig. 1). These faults also have a generally eastward transport direction, but the displacement is estimated as less than that of the Last Chance assemblage, ranging from 0.5 to 5 km (Dunne, 1986). Most of these structures appear to post-date those of the Last Chance assemblage, although the East Sierran and Last Chance assemblages are thought to be broadly coeval and genetically linked (Dunne, 1986).

The Laramide assemblage includes strike-slip faults and west vergent folds that largely post-date Last Chance and East Sierran structures. Although structures of this assemblage have not been identified in the study area, the presence of Laramide structures elsewhere in the southern Inyo Mountains (Dunne, 1986) suggests that some structures in the study area were formed or modified during this period of deformation.

In the study area, evidence of crustal shortening is widespread. East of the crest of the southern Inyo Mountains the structure is dominated by the Cerro Gordo anticline (Merriam, 1963), which involves the entire Paleozoic section in a generally tight, east vergent, and locally overturned fold. The axial trace of this fold, which roughly parallels the northwest trend of the Inyo Mountains, is traceable from near Hunter Canyon, where it is intruded by granitic rocks, to the southern part of the range near Cerro Gordo Peak, where it is more broad and open (Conrad and McKee, 1985). Over most of its length, the plunge of the anticline is nearly horizontal, but near Cerro Gordo Peak the fold plunges about 15° to 20° to the south-southeast. The axial surface of the Cerro Gordo anticline dips steeply westward, so this structure may belong to the East Sierran structural assemblage, as do all other east-vergent folds in the area (Dunne, 1986). Studies of the Permian unconformity by Stone (1984) and Stone and Stevens (1984) suggest, however, that the area underlain by the Cerro Gordo anticline may have been a regional high in the Permian, implying that this fold may have developed during Permian deformation. Later Mesozoic compression may have served to further tighten this fold. If so, an oblique intersection of the Paleozoic and Mesozoic fold belts might explain the variable tightness of Cerro Gordo anticline along its length. Superimposed on this fold are many smaller, mostly northwest-trending, folds and flexures of varying orientation. Some of these folds are East Sierran assemblage structures whereas others are considered to be older folds of the Last Chance assemblage (Dunne, 1986).

In the area of study, Devonian and Mississippian rocks are typically highly foliated, whereas Pennsylvanian rocks are generally unfoliated but are deformed by northwest-trending, tight to isoclinal folds with wavelengths on the order of tens to about 200 m. Late Cenozoic normal faults and associated gravity slumps and landslides obscure many relationships, but palinspastic restoration of Cenozoic offsets suggest that these Paleozoic rocks are part of the axial zone of the Cerro Gordo anticline. In this zone, the Devonian Lost Burro Formation shows substantial thickening by tight, short wavelength folds. Much of this thickening appears to take place at the expense of the Mississippian units, which have been greatly thinned in the axial portion of the anticline; above the Lost Burro Formation, the Perdido Formation and the Tin Mountain Limestone generally are missing and the Rest Spring Shale is only a fraction of the thickness exposed along the crest of the Inyo Mountains. Husk (1979) mapped the Lost Burro-Rest Spring Shale contact as a thrust fault, but actually Devonian rocks lie above Mississippian rocks because the section on the eastern limb of the Cerro Gordo anticline is overturned. Substantial thinning of the shales and thin-bedded limestones of the Mississippian units has occurred, probably by bedding-parallel shear and flow of material away from the axial portion of the Cerro Gordo anticline during folding. Tight folding of the Keeler Canyon Formation in the lower part of San Lucas Canyon and to the east in the Nelson Range suggests that this unit was thickened during formation of the Cerro Gordo anticline.

The Rest Spring Shale commonly is structurally thinned along the trace of the Cerro Gordo anticline throughout the southern Inyo Mountains and actually constitutes a shear zone. Although often only a fraction of the section is present, the unit nonetheless shows a high degree of persistence. Between Keynot Canyon and McElvoy Canyon (Fig. 2), a band of the Rest Spring Shale generally less than 10 m thick extends for 3 km (Conrad and McKee, 1985), whereas 3 km northwest of Cerro Gordo Peak the unit is about 300 m (1,000 ft) thick (Merriam, 1963). This structural thinning may explain the local absence of the lower Mississippian units in many places on the eastern side of the range crest, although some thinning may be a result of forcible intrusion of Mesozoic plutons. In Daisy Canyon, the Rest Spring Shale thins from about 300 m to about 50 m where it bends around a large pluton (Conrad and McKee, 1985). South of Daisy Canyon, the Tin Mountain Limestone and the Perdido Formation are missing and a thin band of Rest Spring Shale is in fault contact with the Lost Burro Formation for about 4 km to a ridge 2 km north of Bonham Canyon, where the fault is intruded by granitic rock. Palinspastic restoration of offset on Cenozoic normal faults to the east suggests that this contact is the same as that between the Lost Burro Formation and Rest Spring Shale exposed 2 km to the east in Sand Canyon (Plate 1), which has been down dropped and repeated.

<u>Cenozoic Structure</u>

Early Cenozoic deformation has not been recognized in the southern Inyo Mountains, although several structures that post-date Mesozoic plutonic rocks could be of this age. These include two west-trending, left-lateral faults south of the area of this study, the Darwin fault and an unnamed fault that cuts across the southern part of Santa Rosa Hills. The Darwin fault was first identified by Kelley (1937) and later studied by Hall and MacKevett (1962) and Stone and others (1989). Approximately 1.6 km of displacement, which mostly pre-dates overlying 5-6 Ma basalts of the Darwin Plateau, is recognized on this fault (Stone and others, 1989). The fault in the southern part of the

Santa Rosa Hills also shows only minor displacement of late Tertiary basalt, but pronounced sinistral drag of adjacent Paleozoic rocks and major structures from the Inyo Mountains to the Darwin Hills suggest about 10 km of left-lateral displacement (Kelley and Stevens, 1975; Stone and others, 1989).

Late Cenozoic structures in the southern Inyo Mountains are characterized by moderate to steeply dipping, northwest-trending normal faults related to basin-and-range extensional tectonics. Uplift along the east side of the Inyo Mountains is at least 3,000 m as shown by the relief from the floor of Saline Valley to the range crest. The dominant structures in the southern Inyo Mountains forming the western and southern boundaries of Saline Valley are the Eastern Inyo fault zone (EIFZ), which is equivalent to the Western Frontal fault zone of Zellmer (1980), and the Hunter Mountain fault zone (HMFZ), referred to by Zellmer (1980) as the Grapevine Canyon fault zone (Fig. 2). Numerous fresh scarps along the range front and in unconsolidated alluvium on the valley floor indicate that these faults are still active (Zellmer, 1980; 1983).

Eastern Inyo fault zone

The Eastern Inyo fault zone (EIFZ) trends approximately N. 30° W. and extends the entire length of the Inyo Mountains. It is readily traceable from Daisy Canyon in the southwestern corner of Saline Valley to Waucoba Wash to the north (Fig. 2) along which it places basin-filling sediment of Saline Valley against Paleozoic and Mesozoic rocks that underlie the steep eastern escarpment of the Inyo Mountains. Zellmer (1980) suggested that this fault has as much as 6,000 m of dip-slip displacement, but this figure seems too high. An estimate here of about 3,900 m of vertical displacement is based on the 3,000 m elevation difference between the floor of Saline Valley and the Inyo Mountains crest plus about 900 m (3,000 ft) of valley fill estimated from gravity studies

(Chapman and others, 1971). In studies of young fault scarps along the EIFZ, Zellmer (1980) noted no evidence of strike-slip displacement.

A dormant extension of the EIFZ south of Daisy Canyon was identified during this study (Plate 1). This extension is suggested by the unbroken southward continuation of the 3,000- to 3,300-m-high crest of the Inyo Mountains and its steep eastern flank to Conglomerate Mesa, approximately 20 km south of Daisy Canyon. The trace of this fault extends from Daisy Canyon, where it brings shattered upper Paleozoic sedimentary rock and Mesozoic granitic rock against a steep escarpment underlain by Mesozoic granitic rocks (Fig. 5), southward to about 2 km north of Bonham Canyon, where the EIFZ is covered by younger, unfaulted Tertiary sediment in Bonham and San Lucas Canyons. From here it is inferred to extend beneath Quaternary sediment southeastward into Lee Flat or Santa Rosa Flat (Fig. 2). The fault gains elevation south of Daisy Canyon and about 3 km north of Bonham Canyon branches into at least 4 faults each dipping 30-50° E. Offset of an overturned contact between the Mississippian Rest Spring Shale and the Devonian Lost Burro Formation indicates a cumulative minimum dip-slip displacement of 430 m on these fault segments.

Hunter Mountain fault zone

The Hunter Mountain fault zone (HMFZ) trends southeastward along the southern boundary of Saline Valley from near Daisy Canyon, where it terminates against the EIFZ (Fig. 2; Plate 1). It extends through Grapevine Canyon (Fig. 2) near Hunter Mountain into Panamint Valley where it merges with the range-bounding fault along the west side of the Panamint Range.

The HMFZ has been described as a "hinge" or "scissors" fault with a pivot point near Grapevine Canyon, with down-to-the-north displacement on the northwestern segment and down-to-the-south displacement on the southwestern segment (Zellmer,

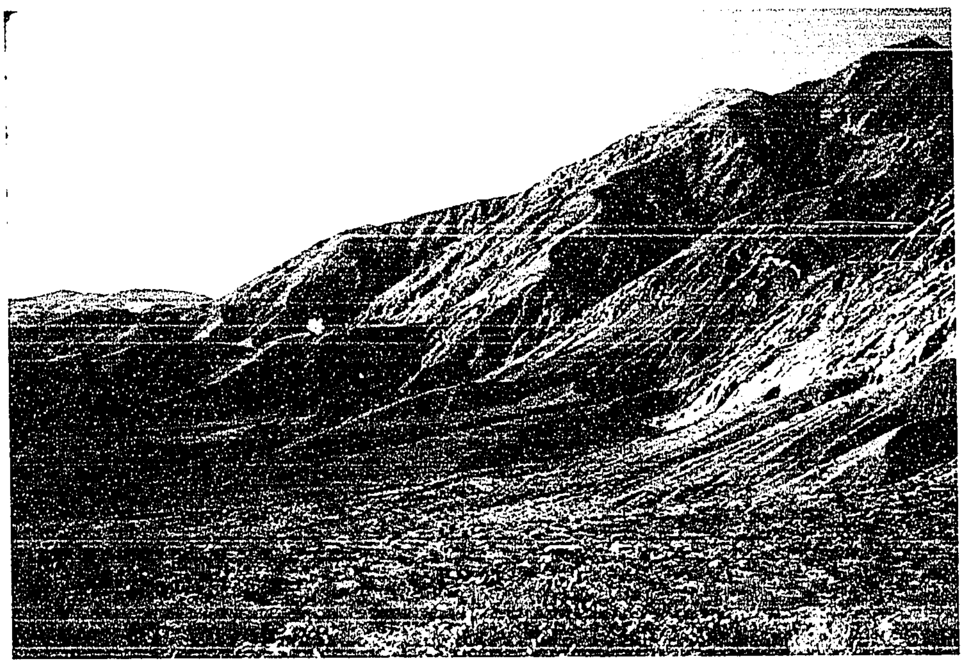

Figure 5. Eastern Inyo fault zone in the southwestern corner of Saline Valley viewed to the south. The large canyon opening into Saline Valley in the center of the photo is Daisy Canyon. The EIFZ extends along the base of the Inyo Mountains on the right-center of the photograph underneath dark-brown perched gravels and small landslide deposits. South of Daisy Canyon, the EIFZ juxtaposes shattered Paleozoic rocks on the extreme left-center against an exhumed footwall of Mesozoic granitic rocks, which underlies the planar surface in the center that dips about 30° to the east. Note the continuity of this surface across Daisy Canyon with footwall rocks on the left. The Hunter Mountain fault zone extends along the range front on the left-center of the photograph and terminates against the EIFZ near the mouth of Daisy Canyon.

1980). Studies by Burchfiel and others (1987) and Sternlof (1988) suggested, however, that much of this apparent dip-slip displacement results from 8-10 km of right-lateral offset. Thus, the south-facing escarpment on the northern edge of Panamint Valley and the steep, north-facing escarpment on the southeastern margin of Saline Valley formed by northwest translation of the Nelson Range relative to Hunter Mountain with little or no dip-slip offset needed to account for the topography. Along the western part of the HMFZ near San Lucas Canyon, both strike-slip and dip-slip displacement of young Quaternary sediment is displayed (Zellmer, 1980), and between Daisy and San Lucas Canyons, older, consolidated talus deposits and basin-fill sediment has been perched along the range front by dip-slip displacement on the HMFZ. The steep escarpment along the HMFZ near Daisy Canyon and the thickness of basin-fill in Saline Valley also indicate substantial dip-slip displacement. A minimum estimate of the total amount of dip-slip offset on the HMFZ near Daisy Canyon is about 2,400 m based on the difference in elevation between the floor of Saline Valley and the Nelson Range (about 1,500 m) plus approximately 900 m of basin-fill in Saline Valley.

Right-lateral displacement along this part of the HMFZ is indicated by the apparent displacement of the large alluvial fans at the mouths of Sand Canyon and San Lucas Canyon (Plate 1). In addition, the mouths of both of these canyons also curve to the east near the range front in a manner suggestive of dextral drag along the HMFZ. Apparent right-lateral displacement along a branch of the HMFZ about 250 m south of the range front has placed older talus deposits about 0.5 km to the east of their likely source, a hill on the east side of the mouth of Sand Canyon (Plate 1), and older gravels at the mouth of Sand Canyon, around which the canyon bends to the east.

Studies by Burchfiel and others (1987) and Sternlof (1988) indicate that the HMFZ serves as a transfer structure that connects Saline Valley and Panamint Valley as paired, pull-apart basins. This interpretation requires that both basins opened at about the same time with a roughly equal magnitude and direction of extension on either side of the HMFZ. The opening of Panamint Valley, which is constrained to have occurred sometime after 4 Ma (Burchfiel and others, 1987), therefore marks not only the inception of movement on the HMFZ but also the down drop of Saline Valley.

Saline Valley was interpreted as a "rhombochasm" by Zellmer (1980) with major range-bounding faults along the four sides of the valley, but the fact that down-dropping of the valley floor has occurred mostly along the western and southern margins of the basin suggests that the basin is actually a west-tilted half-graben. This interpretation is supported by a number of observations. The eastern and northern margins of Saline Valley (Fig. 2) do not appear to be marked by major range-bounding faults. Instead, slopes of the Cottonwood and Saline Ranges are relatively gentle and are overlapped by broad alluvial fans that extend for many kilometers into Saline Valley. In contrast, ranges on the western and southern margins of Saline Valley are marked by the steep escarpments of the EIFZ and the HMFZ, respectively. Along these faults, alluvial fans at the base of the ranges are steep and small, projecting only about 2 km into Saline Valley, and the deepest part of the valley, marked by the Salt Lake playa (Fig. 2), lies only 1.5 km from the Inyo range front. These features indicate that down-dropping of Saline Valley has occurred mostly along the EIFZ and HMFZ and has resulted in westward tilting of the Saline Valley block, with a hinge point in the vicinity of Hunter Mountain.

The Jumble

The formation of modern Saline Valley was accompanied by the development of a tectonic breccia termed the "jumble" by Conrad and McKee (1985). The jumble occurs in a narrow, graben-like feature north of Lee Flat between the Inyo Mountains and the Nelson Range (Plate 1; Fig. 6). The jumble is roughly triangular in shape and covers several square kilometers between the EIFZ, the HMFZ, and San Lucas Canyon (Plate 1).

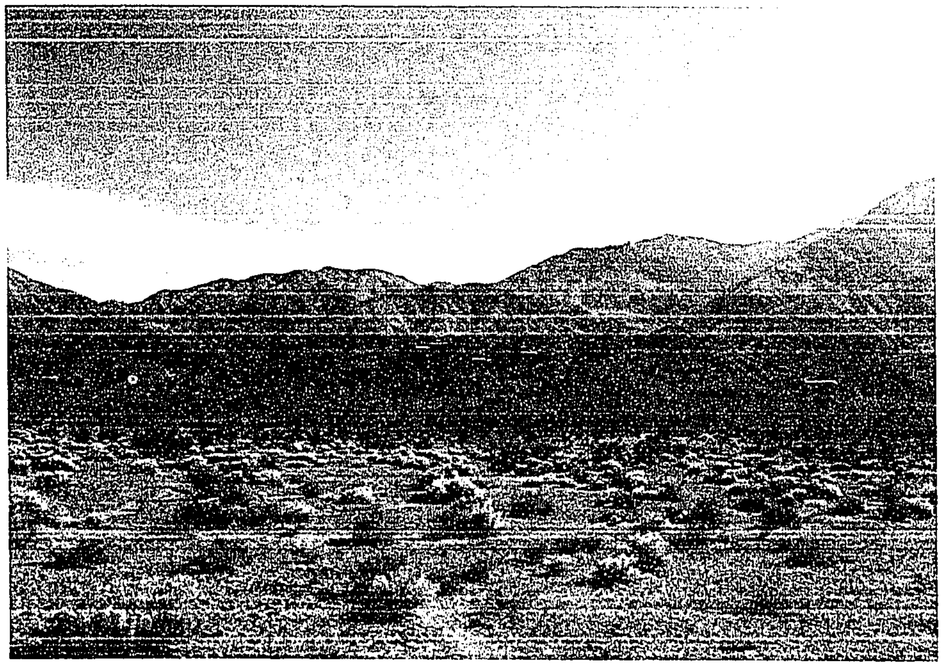

Figure 6. View of the jumble from Saline Valley toward the south. The jumble occupies a graben-like depression between the Inyo Mountains on the right and the Nelson Range on the extreme left of the photo. The Hunter Mountain fault zone extends along the base of the range front from left to right. Jumble deposits form the brown- and tan-colored hills south of the fault that extend to the skyline on the left-center of the photograph.

Viewed from a distance, the topography of the area has a hummocky appearance and is subdued in contrast to more rugged slopes elsewhere in the southern Inyo Mountains.

The bedrock of this area, primarily the Keeler Canyon Formation, is characterized by brittle fracturing and shattering that has broken the rocks into blocks and breccias. The brittle and chaotic nature of the fracturing suggests that the deformation occurred at or near the surface. The size of the blocks varies greatly, ranging from large, relatively intact bodies several hundred meters across, to masses of thoroughly shattered, recemented fragments 1-2 cm across (Fig. 7). From a distance, bedding appears to have some continuity, indicating that individual blocks have not moved far with respect to adjacent blocks, but a general disturbance of the bedding is apparent (Fig. 8). Conrad and McKee (1985) speculated that the brittle fracturing of the jumble resulted from sliding of the mass from the recently uplifted Inyo Mountains, but the actual mechanics of the fracturing and the timing of the deformation remains unclear.

In map view, the jumble is cut into dozens of fault blocks by a system of faults that trend roughly east-west and dip 30-50° to the north (Plate 1). These fault blocks are in turn variously broken and fractured but not greatly displaced or mixed. The fracturing and shattering of the bedrock in general has resulted in poor exposure, so that mapped faults probably represent only a fraction of the total number present, and the sense of offset on the identified faults commonly is difficult to determine. Where discernible, however, offset is down-to-the-north. In map view, most of these faults have a pronounced arcuate shape, convex to the south. This shape results both from the north dip of the faults and the apparent concave-up scoop shape of the fault surfaces, which are reminiscent of slump surfaces. In some places, bedding in the shattered Keeler Canyon Formation is parallel or sub-parallel to the faults so bedding probably at least partly controlled the orientation of the faults.

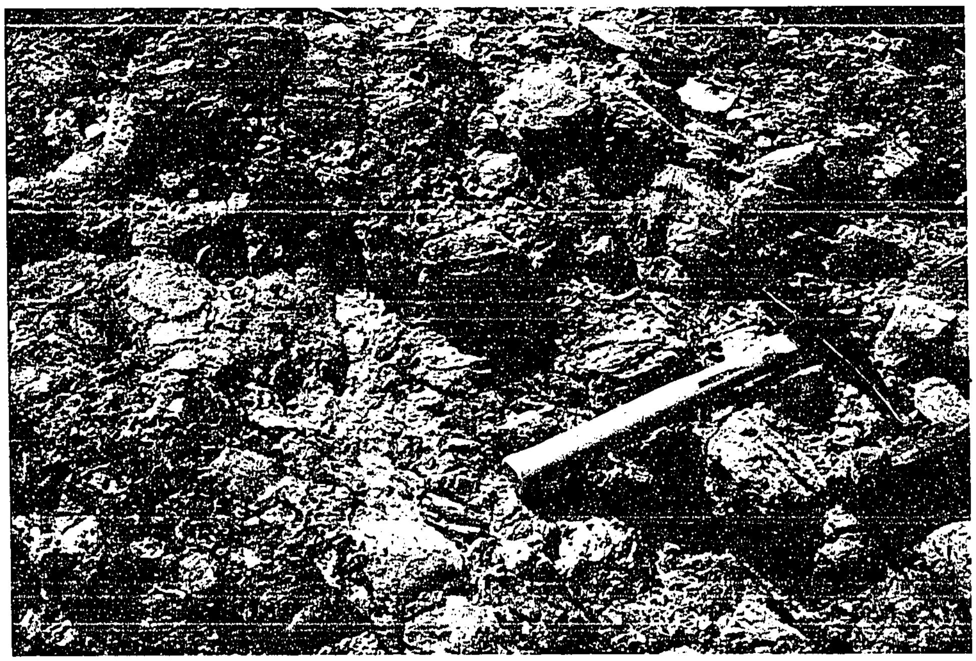

Figure 7. Shattered fragments in the jumble in outcrops on the west side of Sand Canyon. Rock fragments composed of silty limestone of the Keeler Canyon Formation have been recemented, but open cracks suggest continuing deformation.

Figure 8. Eastern side of Sand Canyon showing shattered Keeler Canyon Formation. View is to the east. White and blue banding on the canyon wall shows a general stratigraphic continuity but beds have been disrupted by shattering and numerous, mostly small, faults that strike roughly perpendicular to the canyon wall.

Several landslides of variable size were identified in the area of the jumble, and locally, the distinction between low-angle fault blocks and landslides is not always clear. Near the head of Sand Canyon, an east-trending fault with apparent normal offset of about 125 m can be traced into the headwall of a large landslide mass that covers several km^2 in the upper part of Sand Canyon (Plate 1). This mass is composed of at least two and probably several separate landslides. It seems likely that the incompetence of the bedrock has resulted from previous shattering due to older movement on the EIFZ. Three to four km to the east in the Nelson Range, steeper slopes are underlain by unfractured Keeler Canyon Formation and large landslides are not present.

Evidence for Two Stages of Uplift

A number of geomorphic features and structural relationships in the southern Inyo Mountains suggest that uplift has occurred in at least two stages. The steep escarpments of the Inyo Mountains and the Nelson Range along the EIFZ and HMFZ and offsets of Quaternary sediment in Saline Valley indicate recent deformation in the area, the timing of which is constrained by the post-4 Ma opening of Panamint Valley. The inactive extension of the EIFZ south of the HMFZ, which is sealed by the older gravels of Bonham Canyon, suggests an earlier phase of uplift along the EIFZ.

Evidence of offset on the EIFZ that is significantly older than that on the HMFZ is suggested by the morphology of the eastern Inyo Mountains range front. The steep and apparently young escarpment of the EIFZ, which rises about 1,500 m above the floor of Saline Valley, probably is related genetically to post-4 Ma movement on the HMFZ. This represents only about one-half of the total uplift of the Inyo Mountains block, however, which rises to as high as 3,000 m above Saline Valley. The morphology of the Inyo Mountains block also suggests a two-stage evolution. Throughout the southern Inyo Mountains from Cerro Gordo at the south to as far as north as Paiute Canyon (Fig. 2), the upper parts of major canyons are relatively open and broad and tend to have extensive colluvial cover, streams are relatively straight with constant gradients, and east-trending ridgelines have an average slope of about 10°. Below 1,500-1,800 m, the walls of major canyons become extremely steep and rugged, streams are markedly more sinuous and have abundant waterfalls, and ridgelines continue down into Saline Valley with slopes of 25-30°. Cross-sectional profiles across the Inyo Mountains show this steepening along the eastern margin of the range (Fig. 9). This morphology suggests that the upper part of the range was exposed to erosional modification prior to uplift and exposure of the lower part of the range. The older, dissected gravels that overlie the inactive segment of the EIFZ in the vicinity of Bonham Canyon (Plate 1; Fig. 2) lie at about the same elevation as the top of the steep range-front escarpment, suggesting that they are a small remnant of sediment that was deposited along the length of the Inyo Mountains during the earlier episode of uplift along the EIFZ.

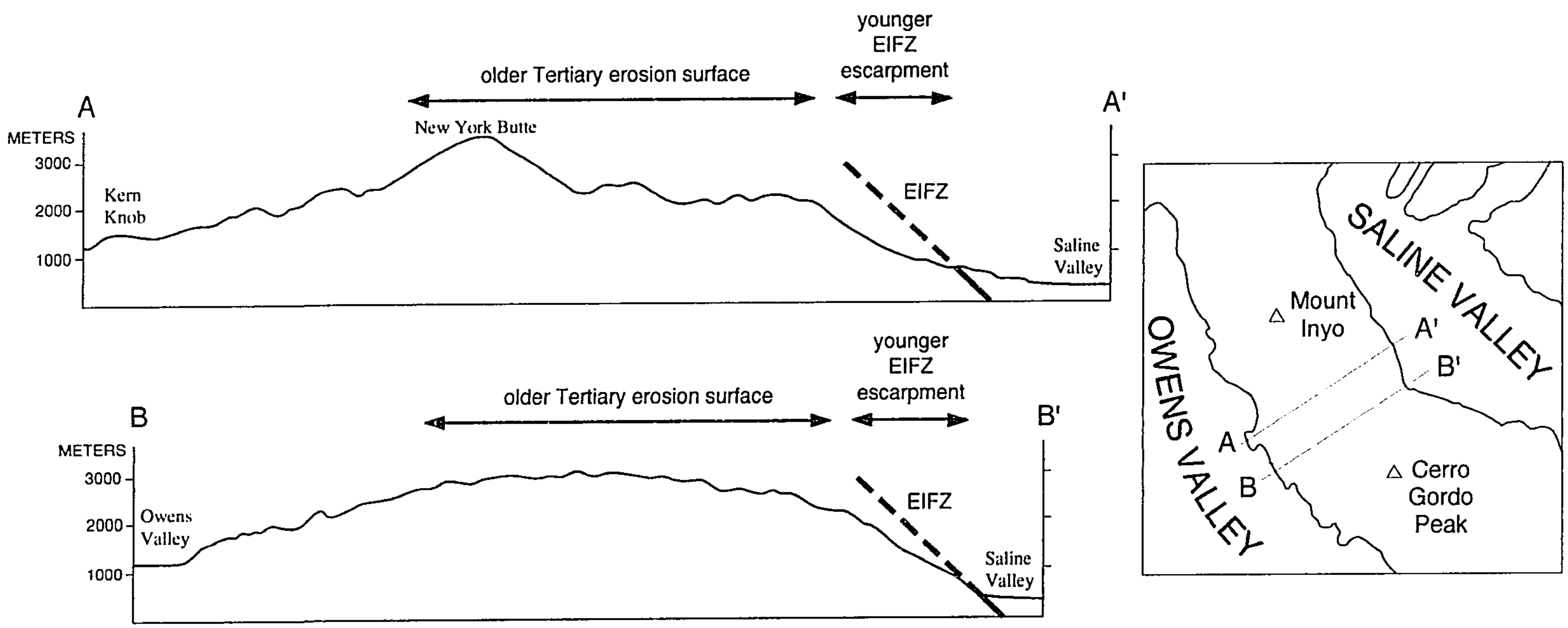

Figure 9. Cross sectional profile of the Inyo Mountains showing the steep, young escarpment of the Eastern Inyo fault zone (EIFZ) on the eastern side of the range and the older erosion surface in the higher elevations.

SEDIMENTOLOGY OF THE GRAVELS OF BONHAM CANYON

A section of older, dissected gravel is exposed along the eastern flank of the Inyo Mountains immediately north of Lee Flat (Fig. 2). This sediment, herein referred to as the gravels of Bonham Canyon, is an important feature for understanding the evolution of the Eastern Inyo fault zone (EIFZ). Its position astride the continuation of the EIFZ suggests that it is an alluvial fan deposit related to an early episode of uplift of the Inyo Mountains block.

Character

The gravels of Bonham Canyon are exposed on the eastern side of the Inyo Mountains east of Cerro Gordo between Bonham Canyon to the north and Lee Flat to the south. The gravels comprise a nearly flat-lying to gently east-dipping, wedge-shaped unit that thins to the east. Although generally poorly exposed, scattered outcrops of the gravels are present in a series of canyons up to 100 m deep that form the upper part of the San Lucas Canyon drainage basin.

The gravels consist almost entirely of moderately- to well-cemented pebble to cobble conglomerate (Fig. 10). Sand- and silt-sized grains are abundant in the matrix, but sandstone and siltstone beds are extremely rare, occurring only as thin interbeds between conglomerate beds. Locally, beds containing boulder-size clasts up to 1 m in diameter are present. The unit appears to fine slightly to the east with both average and maximum clast size decreasing and the amount of matrix increasing eastward. Bedding ranges from indistinct or massive to medium bedded.

Beds interpreted as debris-flow deposits are common. These are typically about 0.5 m thick, matrix-supported, poorly sorted, and contain angular to subangular clasts in ungraded to reverse-graded beds typically with erosional bases. The debris-flow deposits

Figure 10. Outcrop of the gravels of Bonham Canyon.

are interlayered with moderately- to well-sorted, clast-supported beds, 10-30 cm thick, composed of subangular to subrounded clasts in well-developed, fining-upward sequences with moderately- to well-developed horizontal stratification. These beds are interpreted to be stream-flow deposits.

The gravels of Bonham Canyon characteristically weather to smooth, gravel-covered slopes. Consequently, outcrops are infrequent and individual beds only rarely can be followed with confidence. For this reason, only general thickness estimates can be made. The paucity of marker beds makes faults difficult to identify, but one moderately resistant bed was followed with reasonable confidence eastward along the bottom of Bonham Canyon for about 1.5 km with no apparent offset, suggesting little or no displacement by faults in the western two-thirds of the gravel outcrop. Most of the outcrop is characterized by relatively uninterrupted slopes, but in the eastern part of the area, a break in slope along the ridge lines suggests the presence of a small, north-trending fault with down-to-the-east offset of 30-50 m (Plate 1). Seen from a distance, the numerous ridges underlain by the gravels seem to define a single, dissected surface that may be an older pediment (Fig. 11). If so, this pediment surface post-dates significant offset on any faults that may cut the gravels.

Only a minimum estimate of the thickness of the gravels is possible because the base of the unit is exposed only on the thin, eastern side of the gravel exposures. There, the gravels lap unconformably onto rocks of the Pennsylvanian and Permian Keeler Canyon Formation and the Mississippian Rest Spring Shale (Plate 1). Along the western contact in Bonham Canyon where the gravels abut Paleozoic rocks, at least 245 m of section is exposed. Projection of an average eastward dip of about 5° for the gravels adds a minimum of 120 m of section along the lower part of Bonham Canyon, indicating a minimum total thickness of about 365 m.

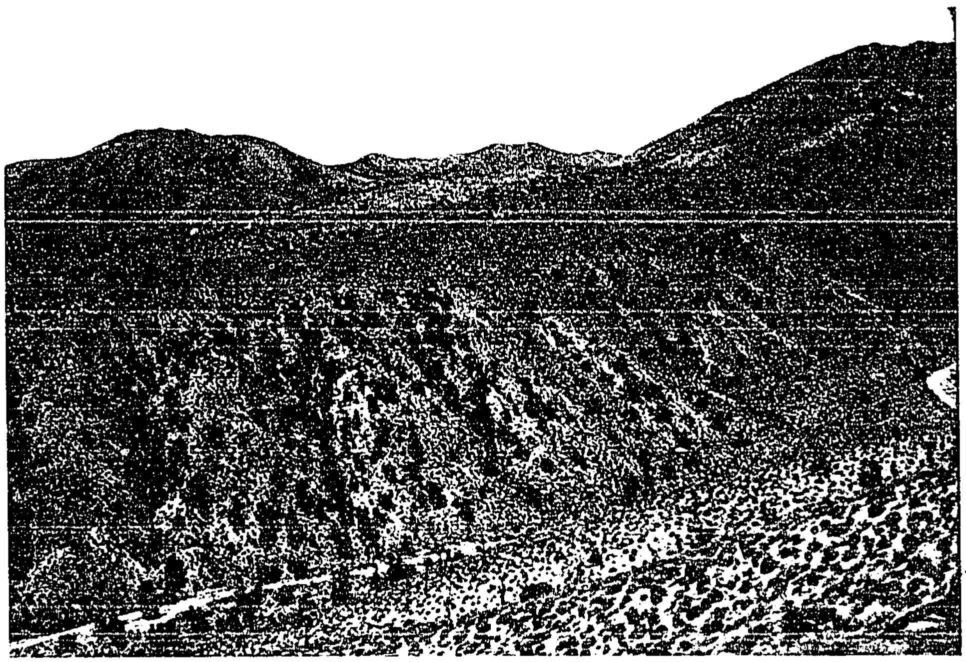

Figure 11. View southward of the pediment surface on top of the gravels of Bonham Canyon. The road in the bottom of the canyon, which is about 70 m deep, provides a scale.

Paleocurrents

A total of 169 measurements of paleocurrent directions in the gravels of Bonham Canyon were measured at 18 localities (Fig. 2) in order to determine possible source areas. B-axis imbrication of gravel clasts was used for current direction and all measurements at each locality were taken from a single bed, except for locality 22 where paleocurrent directions from two beds were measured. Current directions for both beds at locality 22 were calculated separately. Modal directions were established according to a method described by Picard and Andersen (1975).

Paleocurrent directions for the 18 localities are shown on Figure 12 and Table 2. Calculated paleocurrent directions generally are eastward, ranging from N. 10° W. to S. 13° W. When the results are pooled to give an average direction for the entire unit, the vector resultant direction is N. 52° E. with a 95 percent confidence limit of ±7° (Fig. 12).

Provenance

In addition to paleocurrent data, clast composition provides valuable information on possible source areas for the gravels of Bonham Canyon. Because these deposits probably represent material deposited very near-source, relative resistance of the clasts to physical breakdown in a fluvial environment should introduce no significant bias in the results. Clast types were counted at the 18 paleocurrent localities (Fig. 12). A minimum of 100 clasts were counted at each locality and, where possible, all counts were made on a single bed. Efforts were made to count all clasts along a given bed until the desired minimum number was reached; only clasts judged too small for accurate lithologic determination (less than about 3 mm) were ignored. Clast compositions for each locality are shown in Table 3.

Clasts representing almost all lithologic types exposed in and near the southern Inyo Mountains occur in the gravels. Limestone and dolomite are the dominant clast

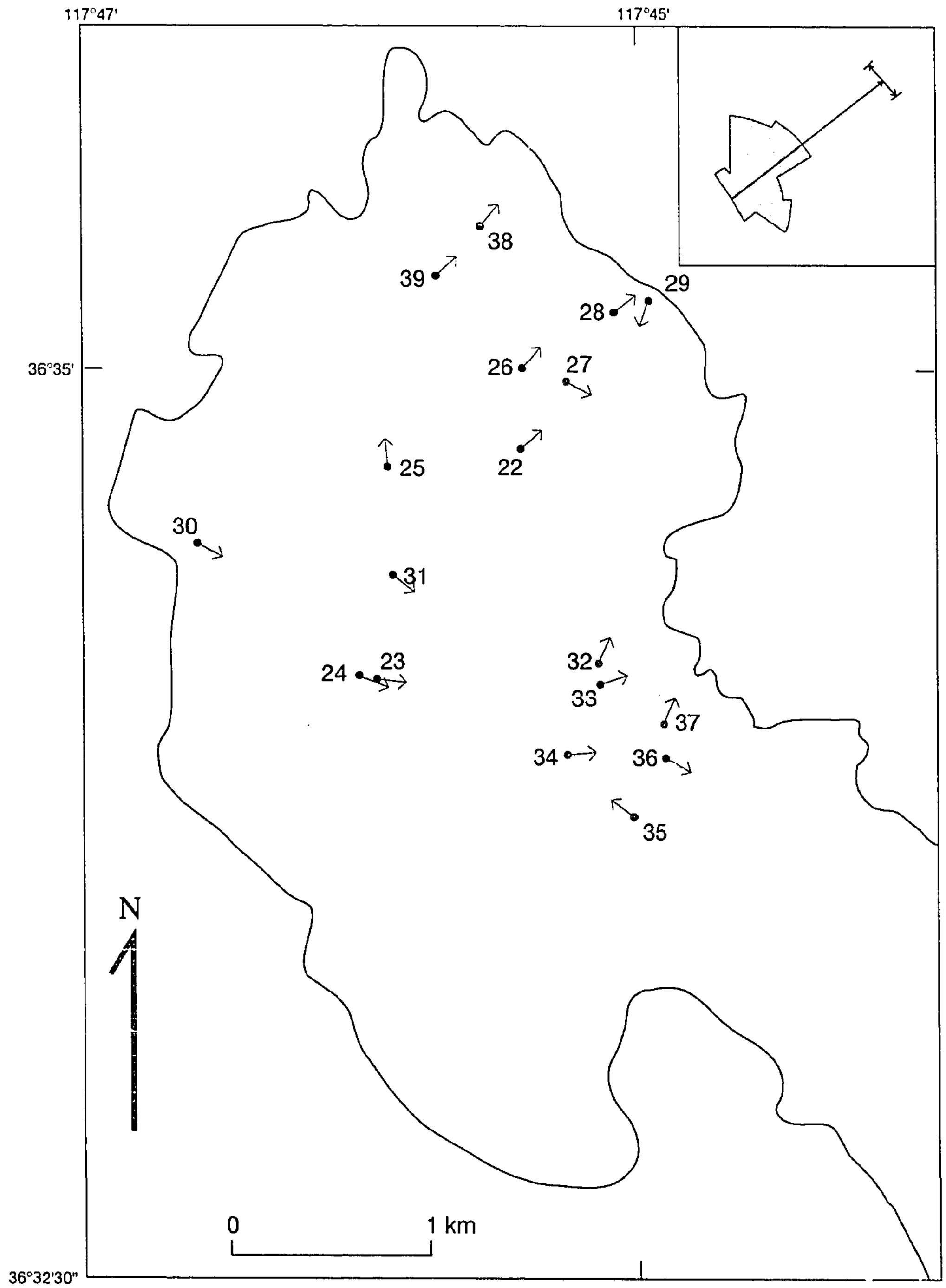

Figure 12. Paleocurrent directions in the gravels of Bonham Canyon. Numbered dots show location of data shown in Tables 2 and 3. Rose diagram shows vector resultant direction and 95 percent confidence limits for entire unit.

Table 2. Paleocurrent directions in the gravels of Bonham Canyon

Locality	Number measured	Resultant (degrees)
22A	4	38
22B	3	64
23	13	101
24	8	103
25	5	8
26	14	38
27	10	16
28	5	57
29	6	193
30	7	118
31	9	12
32	11	19
33	11	80
34	9	85
35	10	350
36	9	119
37	11	22
38	15	38
39	9	41
all data	169	52
magnitude	119.92	
theta	7	

Table 3. Clast composition data for the gravels of Bonham Canyon

LOCATION NUMBER	Clast Lithology	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	39	total	percent
CARBONATE ROCKS	Limestone	76	112	75	78	73	35	21	26	81	20	70	19	26	99	64	45	27	91	1038	46.8
	Dolomite	14	21	39	7	12	43	12	2	13	71	15	81	56	6	45	39	5	4	485	21.9
	Totals	90	133	114	85	85	78	33	28	94	91	85	100	82	105	109	84	32	95	1523	
SILICICLASTIC SEDIMENTARY ROCKS	Conglomerate		1																	1	0.0
	Sandstone		3		1	2	3		1		2	1						15		28	1.3
	Siltstone		6	8	2	8	2	61	76	6	2	7		6	2	19	6	119	27	357	16.1
	Argillite		2		1		5		2			5	2	5			4	20		46	2.1
	Shale			2		4	3			3		2	1			5				20	0.9
	Chert	2	1		3	2					3		2	1	4					18	0.8
	Quartzite			5			2	6			9	1	17	8	1	4	2		5	60	2.7
	Totals	2	13	15	7	16	15	67	79	9	16	16	22	20	7	28	12	154	32	530	
IGNEOUS ROCKS	Igneous							5												5	0.2
	Granite	7	6	13	8	6	9	12	2	8	9	4	6	8	7	10	5	10	4	134	6.0
	Epidote							5												5	0.2
	Pumice								6											6	0.3
	Totals	7	6	13	8	6	9	22	8	8	9	4	6	8	7	10	5	10	4	150	
ALTERED ROCKS	Vein Quartz													1		2		1		4	0.2
	Breccia	2		1		2				3			1				2	1		12	0.5
	Totals	2	0	1	0	2	0	0	0	3	0	0	1	1	0	2	2	2	0	16	
	TOTAL CLASTS	**101**	**152**	**143**	**100**	**109**	**102**	**122**	**115**	**114**	**116**	**105**	**129**	**111**	**119**	**149**	**103**	**198**	**131**	2219	

types in the gravels, but other abundant clast types include sandstone, siltstone, shale, chert, granitic rock, and quartzite. The only major rock type present in the region not found in the gravels is basalt. Limestone and dolomite make up 68 percent of the clasts counted. Siliceous sedimentary rocks are the second most abundant component of the gravels, making up about 24 percent of the clasts. Of this, siltstone is the most abundant rock type, comprising about two-thirds of the total. Quartzite, argillite, sandstone, shale and chert also are important components, but in general they make up less than 3 percent of the gravels. Igneous rocks make up about 7 percent of the total. Granitic rock, consisting mostly of biotite-hornblende quartz monzonite, comprises about 85 percent of the igneous clasts. Other minor igneous rock components include altered intrusive rock, epidote (presumably from small skarns), and pumice. Vein quartz and silicified fault breccia make up about 1 percent of the gravels.

Taken as a whole, clast compositions of the gravels of Bonham Canyon are consistent with the west to southwest source for the gravels indicated by paleocurrent data. The area directly southwest of the gravels comprises the upper part of Bonham and San Lucas Canyons in the Inyo Mountains north of Cerro Gordo Peak (Plate 1). This area is underlain by a generally complete, although faulted, section of middle Ordovician to lower Permian rocks that consists largely of limestone and dolomite with smaller but significant proportions of siliceous rock that includes quartzite, siltstone, and shale (Table 1). This section comprises the upright, western limb of the Cerro Gordo anticline and in general dips 40-60° to the southwest (Plate 1; Fig. 13). Two small quartz monzonite stocks cut the Paleozoic rocks, but make up only a small part of the total drainage area. Certain other areas, however, can be ruled out as source areas on the basis of the overall clast composition. For example, the Nelson Range (Fig. 2) and areas farther east are underlain almost entirely by the Pennsylvanian and Permian Keeler Canyon Formation and rocks of the Hunter Mountain batholith. An eastern source would therefore have

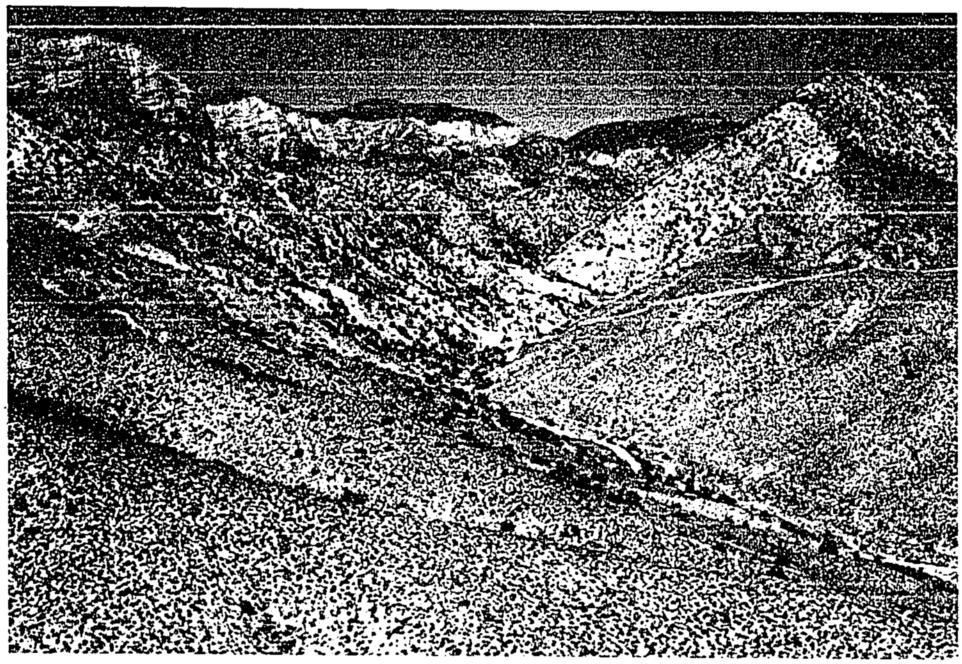

Figure 13. View of the eastern Inyo Mountains showing the interpreted source area for the gravels of Bonham Canyon. Smooth gray slopes on the lower half of the photo are the gravels. White-colored workings in the center are the Bonham Talc Mines, developed in altered dolomite of the Devonian and Silurian Hidden Valley Dolomite. The blue- and white-banded cliffs are underlain by the Devonian Lost Burro Formation, and the dark-blue band near the top of the cliffs is the Mississippian Tin Mountain Limestone. Dark-brown rocks of the Mississippian Perdido Formation and Rest Spring Shale are exposed along the crest.

contributed greater proportions of granitic material and limestone and led to far lower proportions of dolomite than occur in the gravels. Possible source areas in the southern Inyo Mountains south of Cerro Gordo Peak also are relatively poor in dolomite. Areas to the north in the Inyo Mountains are underlain by large masses of granitic rock and therefore also appear to be unlikely source areas for the gravels.

Paleocurrent and clast-composition data for the gravels of Bonham Canyon collectively indicate that the most likely source area is directly to the southwest of the gravels in the southern Inyo Mountains, a hypothesis that can be further tested by comparing variations in clast composition in lower, middle, and upper parts of the gravel section with the bedrock stratigraphy of the southeastern Inyo Mountains. The probable source area for the gravels comprises a relatively undisturbed, upright section that can be roughly characterized, from bottom to top, as consisting of Ordovician and Silurian dolomite and quartzite, Devonian limestone, Mississippian siltstone and shale, and Pennsylvanian and Permian limestone (Table 1). Variations in clast composition upward through the gravel section, therefore, would be expected to record the erosion of the Paleozoic section from top to bottom. This comparison is hampered by the generally poor exposure of the gravels which allows only approximate estimates of stratigraphic position for any given outcrop. Assuming an average dip of about 5° east-northeast for the gravels and projecting counted localities onto a cross section through the gravels trending N. 52° E. from the mouth of Bonham Canyon, an approximate division of the localities into upper, middle, and lower parts of the gravels was made (Fig. 14). Limestone, dolomite, quartzite, and siltstone (including argillite and shale) were selected as clast types likely to reflect the stratigraphic position of the source.

Figure 15 shows relative proportions of limestone, dolomite, quartzite, and siltstone (normalized to 100 percent) from the sampled localities in the gravels of Bonham Canyon. High proportions of siltstone, comprising more than 60 percent of the

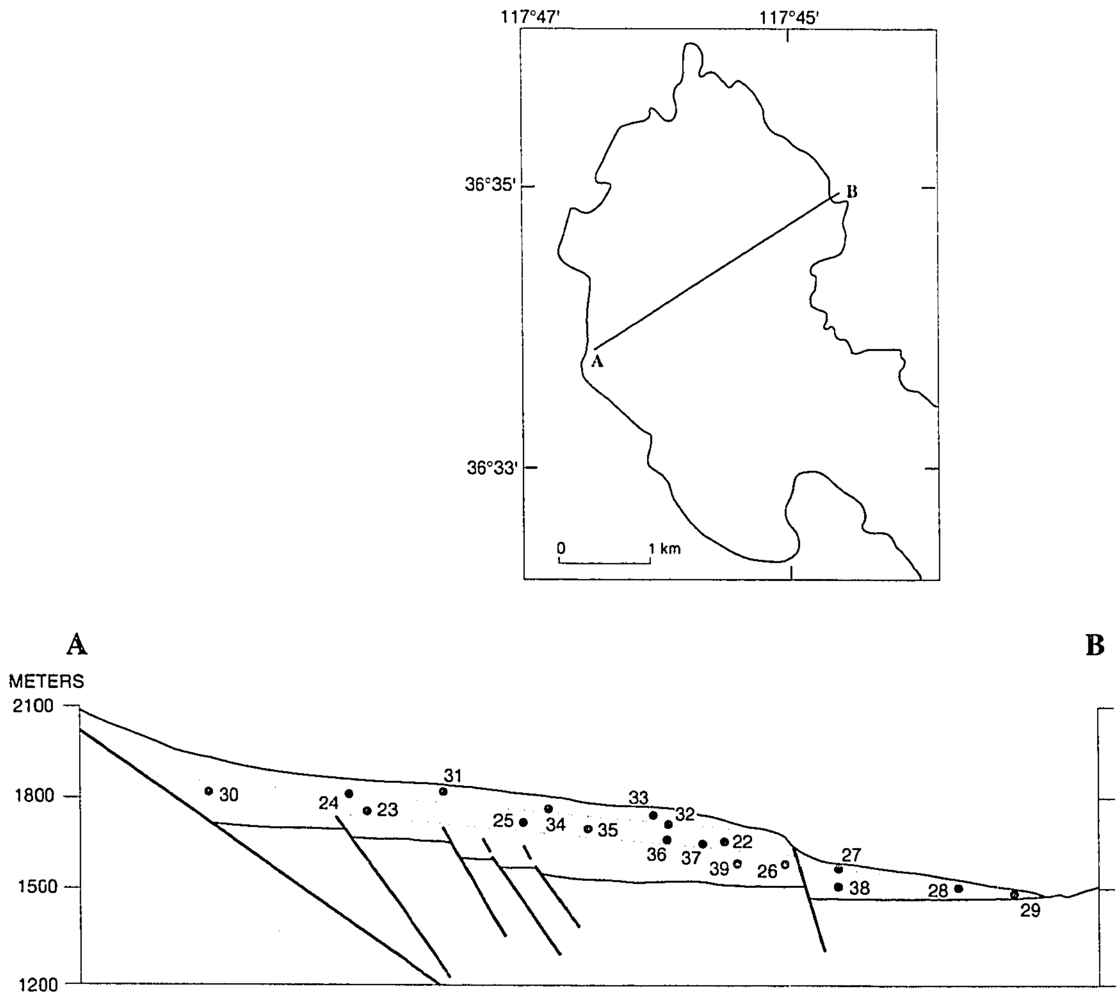

Figure 14. Cross sectional profile of the gravels of Bonham Canyon showing approximate stratigraphic level of clast count localities. Localities were projected onto the cross section using the generalized strike and dip of the beds. See Figure 12 for areal location. Dotted lines show the upper, middle, and lower members based on clast counts and an assumed 5° ENE dip. Heavy lines are faults of the Eastern Inyo fault zone.

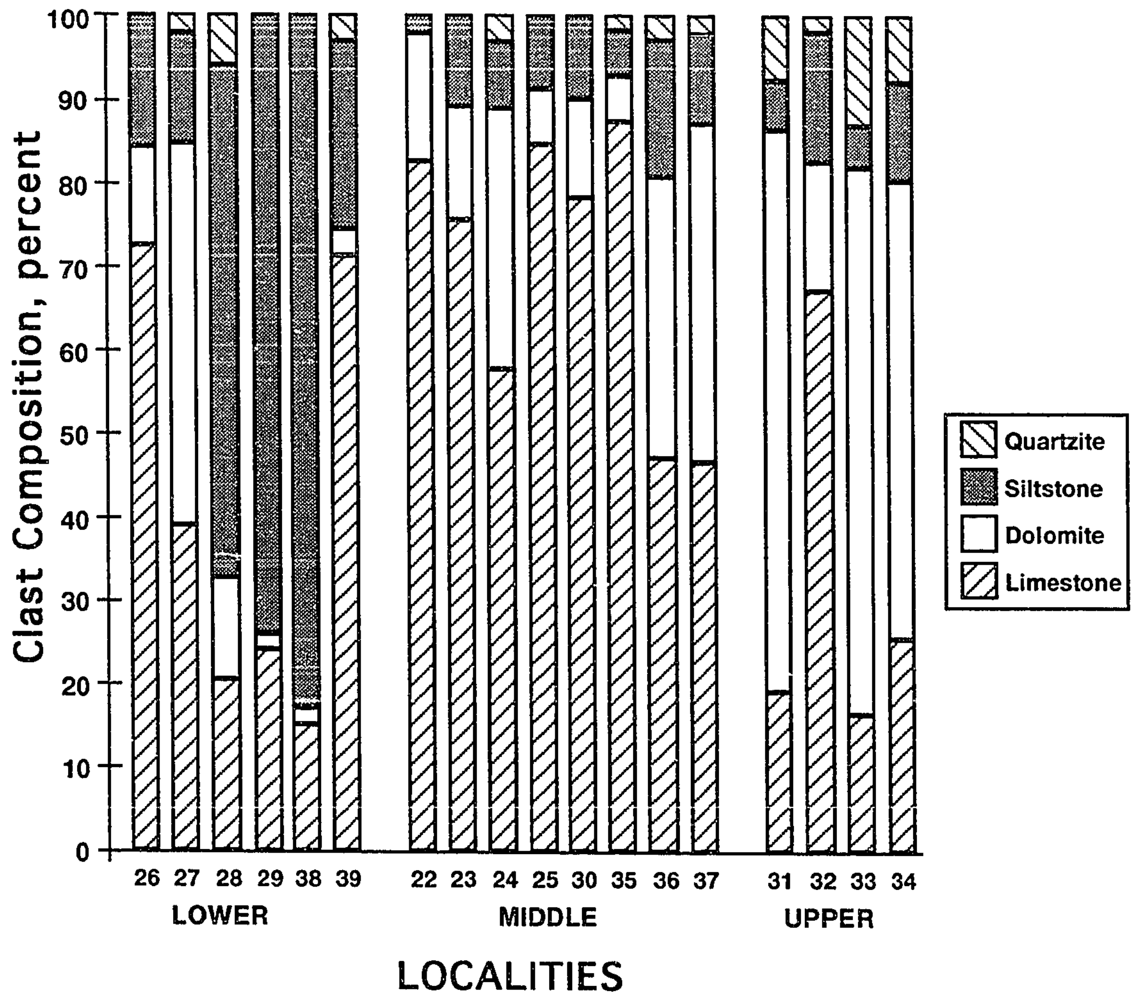

Figure 15. Clast counts normalized to 100% from sampled localities in the gravels of Bonham Canyon, grouped into lower, middle, and upper members by approximate stratigraphic position (see Table 3 and Figure 14).

clasts in 3 of the 6 localities, are apparent in the lowermost sampled beds. Dolomite is rare in these beds, comprising more than about 10 percent of the clasts at only one locality. Quartzite occurs at 3 of the localities, but in amounts less than 5 percent. The abundance of siltstone clasts suggests that the main bedrock source for these beds is the Mississippian Rest Spring Shale.

Localities thought to represent the middle part of the gravel section are characterized by a dominance of limestone clasts. Limestone proportions range from about 45 percent to nearly 85 percent of the clasts, whereas siltstone comprises only about 10 percent of the clasts. Dolomite, although not abundant, generally is more common than in the lower beds, averaging 10-15 percent to as much as 40 percent of the clasts. Minor amounts of quartzite occur at 4 of the 8 localities. The dominance of limestone clasts in these beds suggests that the main source of the clasts is from the Devonian Lost Burro Formation.

In contrast to the lower and middle parts of the gravel sequence, the uppermost sampled beds are characterized by abundant dolomite clasts. Dolomite typically comprises 50-70 percent of the clasts and the ratio of dolomite to limestone is greater than 3:1 in 3 of the 4 localities. Quartzite is present at all 4 localities and comprises at least 8 percent, and up to 15 percent of the clasts in 3 localities. Siltstone comprises less than 15 percent of the clasts. The abundant dolomite clasts and ubiquitous quartzite in these beds indicate a source underlain by the Devonian and Silurian Hidden Valley Dolomite, the Silurian and Ordovician Ely Springs Dolomite, and the Ordovician Eureka Quartzite and Badger Flat Formation.

Thus, the gravels are roughly divisible into three vertically disposed members: 1) a lower Mississippian siltstone-rich member, 2) a middle Devonian limestone-rich member, and 3) an upper Ordovician-Silurian dolomite-rich member. These members record an inverted stratigraphic sequence, and thus unroofing of the source area, and are consistent with the hypothesis of a western source for the gravels in the southern Inyo Mountains.

Age

The gravels of Bonham Canyon were considered by Merriam (1963) and McAllister (1956) to be Quaternary in age. No reason was given for these age designations, and it is likely that the authors considered the gravels lateral equivalents of the basin-filling Quaternary sediments of Lee Flat. McAllister (1956), however, reported conglomeratic sediment underlying olivine basalt less than 3 km east of exposures of the gravels of Bonham Canyon in the northern part of Lee Flat. This olivine basalt is probably part of the Darwin Plateau volcanic field, most of which was extruded between about 6.0 and 5.3 Ma (Schweig, 1982), so the underlying gravels could correlate with the gravels of Bonham Canyon.

No basalt occurs in the gravels of Bonham Canyon either as flows or clasts, and no basalt overlies the main exposure of the gravels. Eroded remnants of flows of the Darwin Plateau volcanic field exposed nearby in Lee Flat, however, suggest that they may have extended over areas now underlain by the gravels of Bonham Canyon. The lack of basalt clasts in the gravels suggests that the gravels are older than any basalt in the vicinity, most of which ranges from 3-6 Ma.

The only age information available is provided by a tephra unit exposed within the gravels near the intersection of Bonham and San Lucas Canyons near locality 29 (Fig. 12) where the gravels lap onto Paleozoic rock (Plate 1). This tephra, located about 5 m stratigraphically above the base of the gravel section, is about 4 m in thickness, and white to buff in color. Subrounded to angular, pebble- to cobble-sized clasts of mostly siltstone and shale with minor limestone and pumice in the tephra indicate that it is water-deposited. The unit is weakly stratified, grading upward from a clast-supported to matrix-supported conglomerate in a light-colored, tuffaceous matrix. Erosional contacts at the base of depositional sequences show cut and fill features. Bronze-colored, but apparently fresh, biotite makes up about 5 percent of the tuffaceous matrix and was used for K-Ar analysis. Analytical procedures are in the Appendix.

K-Ar analysis of biotite from the tephra gave an age of 13.07±0.65 Ma (Table 4), suggesting that deposition of the gravels began in the middle Miocene. Because the tephra was clearly reworked, the possibility that older biotite derived from the Mesozoic granitic rocks was present as a contaminant in the tephra was considered. Biotite from Mesozoic rocks in the vicinity gives ages ranging from about 90 to 160 Ma (McKee and Nash, 1967; Everndon and Kistler, 1970; Griffis, 1986). Although granitic rocks are not a large component in the probable source area for the gravels, they are present and clast counts of beds near the tephra showed up to 10 percent granitic clasts (Table 3). Mesozoic biotite would contain 10 to 30 times more radiogenic argon than biotite of late Cenozoic age, so even a small component of Mesozoic biotite in the tephra would lead to a significant and anomalously older apparent age for the tephra.

The possibility of older, contaminant biotite in the tephra was evaluated by the use of $^{40}Ar/^{39}Ar$ laser-fusion analyses of small populations of biotite grains. The small grain size of biotite did not permit analyses of single grains because insufficient argon was present for accurate measurement, so splits of 3 to 6 grains were used in three separate $^{40}Ar/^{39}Ar$ laser-fusion analyses. Within this small number of grains any variation in age between individual grains would lead to discordant ages, particularly because the likely age of any contaminant grains would be at least 90 Ma. Results of the analyses (Table 4) are concordant within analytical uncertainties and are consistent with the K-Ar age, indicating no contamination of the biotite by older material. Radiogenic argon yield for this sample was consistently low, about 15 percent in all of the K-Ar and $^{40}Ar/^{39}Ar$ analyses, and indicates that the biotite carries a high proportion of adsorbed

Table 4 Results of K-Ar and $^{40}Ar/^{39}Ar$ analyses of biotite from tephra near the base of the gravels of Bonham Canyon

K-Ar Analysis

Experiment No.	K_2O (percent)	$^{40}Ar_{rad}$ ($\times 10^{-11}$ moles/gm)	$^{40}Ar_{rad}$ (percent)	Age$\pm 1\sigma$ (Ma)[1]
871536	7.53	1.4220	15.21	13.07±0.65

$^{40}Ar/^{39}Ar$ analyses [2]

Experiment No.	40/39	38/39 ($\times 10^{-2}$)	37/39 ($\times 10^{-2}$)	36/39 ($\times 10^{-2}$)	$^{40}Ar_{rad}$ ($\times 10^{-14}$ moles)	$^{40}Ar_{rad}/^{39}Ar_K$	$^{40}Ar_{rad}$ (percent)	Age$\pm 1\sigma$ (Ma)[1]
L1024-1	8.6165	1.8036	4.5222	2.4579	4.1853	1.3548	15.72	13.63±0.53
L1024-2	11.462	1.6582	7.2796	3.3987	5.9026	1.4227	12.41	14.31±0.57
L1024-3	10.112	1.8339	5.4685	2.9694	9.0002	1.3394	13.25	13.47±0.50
							Average (K-Ar and $^{40}Ar/^{39}Ar$ analyses)=	**13.62±0.52**

[1]Ages calculated using $\lambda_\varepsilon = 0.581 \times 10^{-10}$ yr^{-1}, $\lambda_\beta = 4.962 \times 10^{-10}$ yr^{-1}, $^{40}K/K_{total} = 1.167 \times 10^{-4}$ mol/mol. Errors are estimates of the standard deviation of analytical precision.
[2]Reactor corrections: $^{40}Ar_K/^{39}Ar_K = 0.0024$, $^{39}Ar_{Ca}/^{37}Ar_{Ca} = 7.2931 \times 10^{-4}$, and $^{36}Ar_{Ca}/^{37}Ar_{Ca} = 2.7511 \times 10^{-4}$. J=0.005597.

atmospheric argon. The relatively low radiogenic yield for this sample results in a somewhat high analytical uncertainty for each analysis of about ±5 percent. The maximum difference in age between the 4 analyses is about 9 percent, indicating that the estimated analytical uncertainties are realistic. The average of the 4 analyses gives a best estimate of the age of the tephra bed of 13.62±0.52 Ma.

Summary and Interpretation

The presence of coarse, pebble- to boulder-size conglomerate that comprises virtually the entire section of the gravels of Bonham Canyon and the abundance of debris-flow deposits indicate a very near source and a high-energy depositional environment typical of alluvial fans. Paleocurrent flow directions in the gravels show a general eastward transport direction of material. Some scatter in flow direction is apparent between measured localities, at least part of which is attributable to normal variation in flow direction that occurs in meandering channels on a fan surface. The modal current direction of N. 52° E., which is approximately perpendicular to the trend of the EIFZ, and the clast composition of the gravels indicate a sediment source in the adjacent Inyo Mountains. The bulk composition of clasts is consistent with rocks exposed in the apparent source area, and the upward change in clast type in the gravels is consistent with a progressive unroofing of the source area. Together, paleocurrent and provenance data demonstrate that the gravels were deposited directly adjacent to their source. K-Ar analysis of biotite from a tephra near the base of the gravels gives an age of approximately 14 Ma, indicating that deposition of the gravels began by middle Miocene time.

LATE CENOZOIC TECTONIC HISTORY

Recognition of the extension of the Eastern Inyo fault zone (EIFZ) south of Saline Valley and the timing of displacement on this part of the fault have important implications for the history of late Cenozoic basin-and-range tectonism in the region. In the Inyo Mountains, two separate and distinct periods of extension are evident, one at about 14 Ma and the other which has continued for the last 4 Ma. The combination of these two extension events is responsible for the modern physiography of the region.

Middle Miocene

The earliest episode of Cenozoic normal faulting in the southern Inyo Mountains region resulted in formation of a basin in which the gravels of Bonham Canyon were deposited (Fig. 16a). Paleocurrent and provenance data for these fanglomerates indicate that they were derived from the Inyo Mountains block, which was uplifted along the EIFZ. Provenance data indicate a unique source area for the sediment that rules out any appreciable strike-slip component on the EIFZ. Because the gravels of Bonham Canyon are little disturbed by later displacement on the EIFZ, a period of erosion and deposition following the initial uplift is suggested. Because there is no evidence of significant topography prior to middle Miocene time (Schweig, 1982), the magnitude of middle Miocene offset can be estimated as approximately 1,225 m, the difference in elevation between the crest of the Inyo Mountains near Cerro Gordo (2,750 m) and the elevation of the base of the gravels of Bonham Canyon (about 1,525 m).

The 13.6±0.5 Ma age from the basal portion of the gravels of Bonham Canyon indicates that displacement along this fault began at this time in the middle Miocene. This faulting presumably resulted in the uplift of the Inyo Mountains block along the entire extent of the EIFZ and deposition of sediment in a newly formed basin to the east

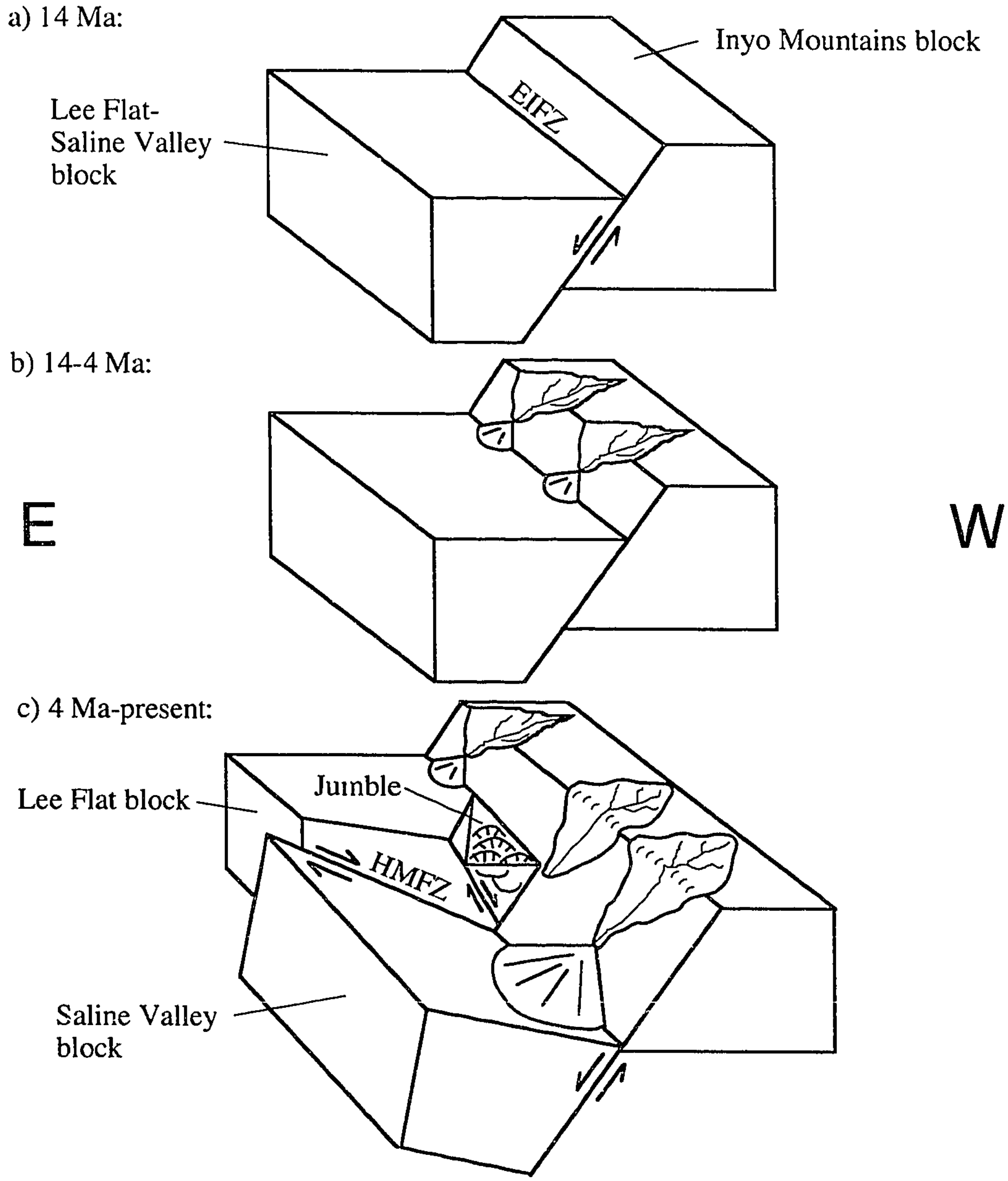

Figure 16. Late Cenozoic structural development of the southern Inyo Mountains as shown by the geology near the southwestern corner of Saline Valley as viewed from the north. a) Movement on the EIFZ begins at about 14 Ma, separating the Inyo Mountains structural block from the undivided Lee Flat-Saline Valley block. b) The uplifted Inyo Mountains block is eroded and the gravels of Bonham Canyon are deposited. c) Renewed extension at 4 Ma forms the HMFZ which separates Saline Valley from the Lee Flat structural block, which is now linked to the Inyo block. Down-drop of the Saline Valley block is accommodated by dip-slip movement along the northern portion of the EIFZ and oblique dextral-normal slip on the HMFZ. The jumble is forming south of the intersection of these faults as the support provided by the Saline Valley block is removed.

that extended the length of Saline Valley and continued as far south as the Darwin Plateau (Fig. 2). The gravels of Bonham Canyon are the last surviving remnants of the sediment deposited in this basin. The gradual decrease in elevation of the crest of the southernmost Inyo Mountains suggests that offset on this fault decreased to the south and it may have died out in the vicinity of the Darwin Plateau. Alternatively, the southern part of this block may have been dismembered by later faulting. For example, an east-trending fault in the southern part of the Santa Rosa Hills, which appears to have about 10 km of left-lateral displacement (Stone and others, 1989), may have cut off the southern portion of the Inyo block. If movement on this fault post-dated initial displacement on the EIFZ, the Darwin Hills and the Argus Range may represent a southern extension of the Inyo Mountains, with the EIFZ extending along the east side of the Argus Range.

The EIFZ appears to have formed as a dip-slip, normal fault at about 14 Ma and was responsible for uplift of the Inyo Mountains block. This event represents a significant period of block faulting in the middle Miocene. Offset along the EIFZ appears to be entirely dip-slip, and the N. 32°W. trend of the EIFZ is interpreted to indicate an extension direction oriented ENE-WSW.

Late Miocene

Following the middle Miocene uplift, the remainder of Miocene time was characterized by tectonic quiescence in the Inyo Mountains (Fig. 16b). During this time, sediment continued to accumulate in the basin east of the uplifted Inyo Mountains block and erosion resulted in development of the mature Tertiary erosion surface still present in the higher parts of the range above the elevation at which streams increase in gradient and canyons become narrow and rugged. This occurs consistently at an elevation of about 1,600 m, such as in Daisy, Craig, Hunter, McElvoy, and Keynot Canyons (Fig. 2) as well as in other minor canyons along the length of the southern Inyo Mountains.

The break in slope that marks the faulted margin of this erosion surface most likely coincides with the former level of the middle Miocene basin represented by the gravels of Bonham Canyon, as these gravels are exposed between about 1,500 and 2,100 m. This relationship also supports the inference that uplift occurred along the entire length of the EIFZ. Total relief on the late Miocene erosion surface ranges from present elevations of about 1,600 m to the 3,000-3350 m crest of the Inyo Mountains, suggesting that the middle Miocene uplift totaled at least 1,400 m. This estimate compares with the minimum estimate of offset of 430 m on several branches of the EIFZ south of Bonham Canyon and 1,225 m of offset indicated by the difference in elevation of the Inyo crest and the base of the gravels of Bonham Canyon in the adjacent basin. The differences between these estimates suggest that there may have been additional movement on other branches of the fault zone or that there already was some relief prior to the middle Miocene.

Pliocene and Quaternary

Pliocene and Quaternary time was marked by renewed tectonism in the southern Inyo Mountains. This tectonism is manifested by renewed dip-slip offset along the EIFZ adjacent to Saline Valley and by a combination of right-lateral and dip-slip displacement along the Hunter Mountain fault zone (Fig. 16c). Down-dropping of the Saline Valley block at 4 Ma probably rejuvenated streams in Bonham and San Lucas Canyons that began to remove the gravels of Bonham Canyon. Continuing tectonic activity in this area is suggested by the rugged scarp north of Daisy Canyon and the HMFZ scarp along the southern boundary of Saline Valley, although only one earthquake of magnitude 3.0-4.4 has occurred in the southwestern part of Saline Valley between 1808 and 1987 (Groter, 1988).

Pliocene-Recent faulting along the EIFZ took place mostly north of Daisy Canyon. Total offset north of Daisy Canyon is at least 2,100 m as estimated by the height of the 1,200 m range front escarpment along the Inyo Mountains plus approximately 900 m of valley-filling sediment in Saline Valley (Chapman and others, 1971). Some of this sediment, however, probably is buried lateral equivalents of the gravels of Bonham Canyon and therefore is middle to late Miocene in age. The younger EIFZ escarpment continues several kilometers south of Daisy Canyon with a maximum apparent offset of about 1,200 m. As the trace of the EIFZ gains elevation farther to the south, apparent offset decreases to near zero about 8 km south of Daisy Canyon where the fault is covered by the gravels of Bonham Canyon, which apparently are undisturbed by post-middle Miocene displacement along this fault.

Detailed studies and interpretation of the HMFZ by Burchfiel and others (1987) in Panamint Valley provide important information relating to the second phase of uplift in the southern Inyo Mountains. These studies indicate that the HMFZ has 8-10 km of right-lateral offset in the vicinity of Grapevine Canyon, based on the offset of the contact of the Hunter Mountain Quartz Monzonite with Paleozoic rocks. This offset accommodated the opening of Panamint Valley on the south side of the fault. The correlation of basalts dated at about 4 Ma on the east and west sides of Panamint Valley (Burchfiel and others, 1987) indicates that this opening, and hence, offset along the HMFZ, took place after 4 Ma. Geophysical surveys (MIT 1985 Field Geophysics Course and Biehler, 1987) and drill-hole evidence (Smith and Pratt, 1957) indicate that basalt is not present under the valley-filling sediment in Panamint Valley, implying that the floor of the valley is not a down-dropped graben. Instead, Burchfiel and others (1987) suggested that Panamint Valley opened by displacement along a shallow, low-angle normal fault dipping about 15° to the west that surfaces along the eastern margin of the valley. This fault projects underneath the Darwin Plateau and other areas to the west, including the southern Inyo Mountains, suggesting that this entire area has moved northwestward relative to the Panamint Range. The HMFZ forms the northern boundary of the low-angle normal fault.

In order to maintain strain compatibility, extension across the Panamint Valley-Darwin Plateau region must be balanced by a roughly equal amount of extension north of the HMFZ (Burchfiel and others, 1987). In the Saline Range (Fig. 2), post-4 Ma extension is accommodated by a series of north-northeast striking, mostly down-to-the-west dip-slip faults with cumulative extension of 8.9±2.5 km (Sternlof, 1988). In Saline Valley, however, this extension is accommodated differently, as faulting resumed on the EIFZ north of the HMFZ by down-to-the-east dip-slip along the western boundary of Saline Valley. Because most of this extension was accommodated on the EIFZ, modern Saline Valley appears to have developed as a half graben, with the valley floor tilted to the west. Offset on the HMFZ along the southern boundary of Saline Valley is therefore oblique (dextral-normal) slip as the valley floor drops down and away from the Inyo Mountains along the east-dipping EIFZ, parallel to the trend of the HMFZ. The differences across the HMFZ led Burchfiel and others (1987) to interpret the HMFZ as a transfer structure that accommodates contrasting styles of extension.

The inception of movement on the Hunter Mountain fault zone marked an important change in the tectonic regime of the southern Inyo Mountains. Whereas renewed extension resulted in a resumption of normal faulting along the EIFZ north of the HMFZ, down-faulting south of the HMFZ was limited to thc extreme northwestern part of the Lee Flat block between the Nelson Range and the Inyo Mountains. The remainder of the Lee Flat block, including the Nelson Range, Lee Flat, and the area underlain by the gravels of Bonham Canyon, comprises a broad plateau with an average elevation of about 1,800 m. This elevation is consistent with the inferred elevation of the basin formed by middle Miocene uplift, indicating that this block has remained in a constant position relative to the Inyo Mountains during the last 4 million years.

The opening of Panamint Valley after about 4 Ma, therefore, probably was coeval with resumed normal faulting along the western margin of Saline Valley reflecting similar amounts of total extension both north and south of the HMFZ. The N. 60° W. trend of the HMFZ is parallel to the direction of extension, approximately WNW-ESE at this time. This extension direction suggests a change from the ENE-WSW middle Miocene extension direction of about 60°. South of the HMFZ, the Inyo and Lee Flat blocks remained coupled, with most of the extension accommodated along the west-dipping, low-angle fault projecting under Panamint Valley. North of the HMFZ, extension was accommodated principally by oblique dextral-normal slip on the HMFZ and dip-slip displacement on the pre-existing EIFZ, forming modern Saline Valley. The HMFZ formed as a link between the pre-existing zone of weakness along the EIFZ and the west side of the Panamint Range.

The recent, rapid down drop of Saline Valley is thought to have led to the formation of the jumble in a roughly triangular-shaped area directly southeast of the intersection of the EIFZ and the HMFZ (Plate 1). The increase in post-middle Miocene offset along the EIFZ from near zero to about 1,200 m from south to north along the western side of the jumble suggests that this area was down-dropped to the north. The chaotic fracturing of the jumble resulted from large-scale mass movement of rock north toward Saline Valley in a series of slides and slumps. This movement is attributed to removal of support following down-dropping of the Saline Valley block producing a hole and leading to a tendency of this area to slump and slide northward (Fig. 16c). Right-lateral drag along the south side of the HMFZ also may have moved the Nelson Range to the east, resulting in an area of tensional strain where the graben-like depression of the jumble lies. The jumble is about 7 km wide along the range front, close to the estimated

8-10 km of right-lateral offset on the HMFZ. Thus, the jumble may have widened progressively with time as the oblique right-lateral slip along the HMFZ pulled the Saline Valley block away from the Inyo Mountains to the east-southeast. In addition, slumping may have been facilitated by fracturing induced by earlier middle Miocene offset along the EIFZ that effectively softened the bedrock of the jumble, and brought relatively incompetent Paleozoic rock in the hanging wall against a footwall of granitic rock.

Summary

The Late Cenozoic tectonic history of the southern Inyo Mountains is characterized by two separate and distinct episodes of extension. The first episode began about 14 Ma with the uplift of the Inyo Mountains block along the EIFZ. This episode marks the initial uplift of the range and represents the beginning of basin-and-range extension in the Inyo Mountains region. Displacement along the EIFZ totaled in excess of 1,225 m and resulted in uplift of the entire Inyo Mountains block. The north-northwest trend of the Inyo Mountains block and the EIFZ and the absence of any strike-slip component on the EIFZ indicates that the extension direction was ENE-WSW during middle Miocene time. The down-dropped eastern block formed a basin that included the western part of Saline Valley and extended south to include the Lee Flat and Darwin Plateau areas. These plateaus are relics of this middle Miocene basin and their position has remained undisturbed relative to the Inyo Mountains block since that time.

Relative tectonic quiet ensued through the remainder of the middle and late Miocene and into the Pliocene. This period was marked by erosion of the uplifted Inyo block and the deposition of coarse fanglomerates on the down-faulted block to the east. The gravels of Bonham Canyon are a small remnant of these deposits that once presumably extended the length of the Inyo Mountains. This erosional period led to the development of the mature late Miocene erosion surface in the upper part of the Inyo Mountains above an elevation of about 1,600 m.

The opening of Panamint Valley soon after 4 Ma signaled a period of renewed extension in the region. This period is characterized by development of the oblique right-lateral HMFZ and resumption of normal faulting on the northern part of the EIFZ. The trend of the HMFZ indicates a change of about 60° in the regional extension direction to roughly ESE-WNW. Apparently, displacement along the southern portion of the EIFZ was incompatible with this extension direction and extension was accommodated to the east along a west-dipping, low-angle normal fault in Panamint Valley. North of the HMFZ, however, extension was accommodated by oblique slip on the HMFZ and dip-slip offset on the segment of the EIFZ north of the HMFZ. These offsets resulted in a westward tilt of the down-dropped Saline Valley block, forming the modern Saline Valley half graben. The jumble developed in the northwestern part of the Lee Flat block as a result of slumping.

REGIONAL IMPLICATIONS

The recognition of two distinct extensional events in the southern Inyo Mountains provides new information regarding the late Cenozoic tectonic evolution of the Inyo-Death Valley region. Although more than one episode of extension in this region has been previously recognized, studies in the Inyo Mountains provide a more complete picture regarding the timing of these events and the orientation of the stress fields that have controlled development of the modern physiography of the region.

Timing of Basin and Range Extension

The 14-Ma age of the gravels of Bonham Canyon, which cover a now inactive extension of the EIFZ, demonstrate that normal faulting began in the Inyo Mountains in the middle Miocene. Some workers (Wright and others, 1984; Schweig, 1982; 1989) have suggested that extension across the Inyo-Death Valley region progressed westward with time. This progression is represented by 14-Ma extension in the Kingston Range, 8 Ma and younger extension in the Black and Greenwater Ranges, volcanism and faulting beginning at about 6 Ma in the Coso Range (Fig. 1), and tectonism beginning at about 4 Ma in the Saline Range and Owens Valley (Fig. 2) (Wright and others, 1984). The new age of faulting in the Inyo Mountains, however, suggests that extension may have begun synchronously throughout the region. The beginning of major crustal extension in Death Valley was marked by the start of deposition of the Artist Drive Formation at about 14 Ma (Cemen and others, 1985). McKenna and Hodges (1990) noted two periods of extension in the Death Valley region: an earlier period lasting from about 15 to 10 Ma, and a later period lasting from 10 to 0 Ma. The new data on faulting in the Inyo Mountains suggest that initiation of extension in the Inyo Mountains and in Death Valley was similar and that younger tectonic events are part of a longer history of extension in the region.

Overall, the data suggest that extension throughout the region occurred in two discrete, major episodes beginning about 14 Ma. The episodic nature of extension in the region is well illustrated in the southern Inyo Mountains where two faulting events separated by about 10 million years are recognized. This episodicity is reflected in other areas as well. In Death Valley, the period following the initial formation of the Furnace Creek basin at 14 Ma is characterized by erosion and deposition of the Artist Drive and Furnace Creek Formations until about 5 Ma (Cemen and others, 1985) when the modern Death Valley basin began to form. In the Panamint Range, an older period of extension is represented by deposition of fanglomerates of the Nova Formation. These fanglomerates, which are overlain with angular unconformity by basalt dated at 5.13±0.35 Ma (Hall, 1971), may be substantially older than the basalt. Schweig (1989) suggested that these deposits may be older than 7.7 Ma. It is possible that the oldest part of the unit is as old as 14 Ma and that these rocks reflect the same uplift history as the gravels of Bonham Canyon and the Artist Drive Formation. Dismemberment of the Nova basin by the post-4 Ma opening of Panamint Valley appears to mark a second major phase of extension.

Reorientation of the Extension Direction

A change in the extension direction from WSW-ENE to WNW-ESE has been recognized elsewhere in the Great Basin (Zoback and others, 1981). Although timing of this change is not well constrained, data from central Nevada led Zoback and Thompson (1978) to suggest that it took place between 14 and 6 Ma. In the Darwin Plateau, Schweig (1982; 1989) also recognized a similar change in extension direction but suggested that it occurred after about 5.7 Ma.

Initial uplift of the Inyo Mountains occurred about 14 Ma along the N. 30° W.-trending EIFZ. Paleocurrent and provenance data for the gravels of Bonham Canyon indicate that there was no strike-slip displacement along the EIFZ during this episode of faulting and extension was thus in a WSW-ENE direction. In Death Valley, relationships are less clear owing to the magnitude of later extension and the strike-slip movement on the Furnace Creek fault zone that has dismembered older basins. Several observations suggest that early extension in the Death Valley region was initially oriented WSW-ENE. The oldest Cenozoic sedimentary deposits, which include the approximately 25 to 20 Ma section at Bat Mountain and the 14 to 6 Ma Artist Drive Formation, appear to have southwestern sources (Cemen and others, 1985). If sediment was carried perpendicular to the uplift, the faults (possibly including the northwest-trending Furnace Creek fault zone) would have been northwest-trending, implying a WSW-ENE orientation of the extension direction.

A second period of extension in the Inyo Mountains, this time oriented WNW-ESE, is indicated by the development of the HMFZ at about 4 Ma. In the Saline Range, extension also reflecting a WNW-ESE direction, is accommodated along northeast-trending, mostly down-to-the-west normal faults that cut 3-5 Ma basalts (Burchfiel and others, 1987). In Death Valley, right-lateral slip on the Furnace Creek fault zone is interpreted to accommodate northwest-directed extension on the southwest side of the fault (Wright and Troxel, 1970; Stewart, 1983).

Several workers have noted that faults having originated during the early period of extension in the Inyo-Death Valley region appear to show variable extension directions. For example, in Death Valley, faults that formed prior to 10 Ma have a variable sense of offset, whereas younger faults show a consistent northwestern extension direction (McKenna and Hodges, 1990). In the Coso Range, older, northwest-trending faults have a variable sense of offset that includes dip-slip, strike-slip, and oblique-slip (von Huene, 1960; Walter and Weaver, 1980; Duffield and Roquemore, 1988). Faults younger than 6 Ma are mostly north-to northeast-trending normal faults (Duffield and Roquemore, 1988). Evidence for a major reorientation of the stress field between 14 and 4 Ma suggests that the apparent variability of displacement on pre-10 Ma faults may be attributed to later (<10 Ma) oblique- or strike-slip displacement on faults that initially formed in response to southwest-directed extension.

Physiography of Extension

Analysis of Cenozoic faulting in the Inyo-Death Valley region allows a reinterpretation of the way in which extension has been accommodated. In addition to new evidence regarding the age of extensional tectonics in the western part of the region, apparently anomalous trends of various structural features can be explained. For example, the present northwest trend of the White-Inyo Mountains block seems anomalous in light of recent northwest-directed extension. Some workers have explained this trend as resulting from a right-lateral component of shear along northwest-trending faults in the region, so that features such as Saline Valley are interpreted as a type of transtensional basin (Lombardi, 1964; Wright, 1976; Zellmer, 1980). Although right-lateral shear can be documented on some faults, such as the Owens Valley fault, the total amount of late Cenozoic dextral offset across Owens Valley is probably negligible (Stewart, 1988). With the exception of the Furnace Creek fault zone, right-lateral offset is not generally apparent on other northwest-trending faults in the region. A better explanation for these northwest-trending structures is that they are inherited from an older period of extension upon which a younger period of extension with a different extension direction has been superimposed.

Thus, it appears that initial development of almost all northwest-trending structures in the Inyo-Death Valley region began in middle Miocene time. Included are the EIFZ and the Furnace Creek fault zone, and the uplifted Inyo and Panamint-Cottonwood blocks. The modern northwest trend of Panamint Valley, therefore, may result from earlier offset along a northwest-trending, high-angle range-bounding fault on the west side of the Panamint Range that served as a break-away zone for the low-angle fault that formed modern Panamint Valley. In the Cottonwood Mountains, which form the northern part of the Panamint block, younger northeast-trending normal faults have dismembered the older western range-front fault, translating the Dry Mountains and the Saline Range to the northwest. On the east side of the range in Death Valley, however, the relict northwest-trend of this part of the Panamint block is preserved with only slight modification by younger faulting (Fig. 1).

The only northwest-trending structure in the region that does not reflect WNW-ESE-oriented extension is the HMFZ. Important differences distinguish this fault from the older northwest-trending structures. This fault strikes about 30° farther to the west and, most importantly, it is primarily a strike-slip fault. It is oriented more or less perpendicular to the northeast-trending normal faults that cut the Saline and Cottonwood Ranges to the north and therefore is probably related solely to WNW-ESE-directed extension.

The two stages of extension in the Inyo-Death Valley region also are apparent in the morphology of the Inyo Mountains. The eroded upland of the Inyo Mountains represents erosion of the range following middle Miocene uplift, and remnants of old and possibly correlative erosion surfaces occur throughout the region in the higher elevations (Fig. 17). Maxson (1950) noted such an erosion surface in the upper part of the Panamint Range, and a similar surface, obscured by younger faulting in most places, is present in the vicinity of Tin Mountain and Dry Mountain in the Cottonwood Range (Fig. 17). An old erosion surface, which consists of a broad plateau at an elevation of about 2,100 m surrounded by young basins 600 to 1,500 m lower, is particularly well preserved around

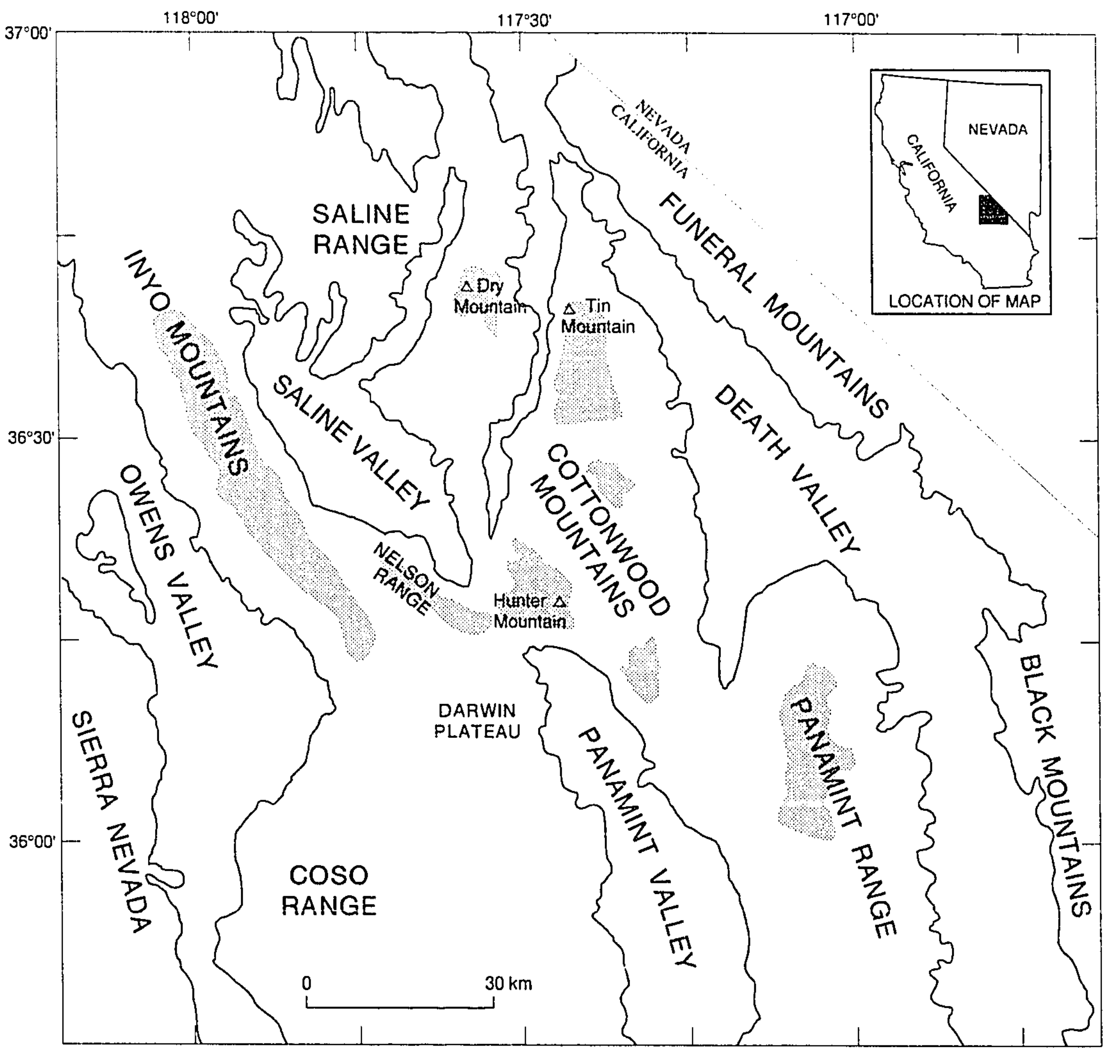

Figure 17. Relict Tertiary erosion surfaces in the Inyo-Death Valley region.

Hunter Mountain. Pliocene and Quaternary faulting has resulted in the modern rugged range fronts and stranded these older surfaces in the upper parts of the ranges where rejuvenated streams are beginning to dissect them.

CONCLUSIONS

Study of the structure, geomorphology, and Cenozoic sedimentology of the eastern side of the Inyo Mountains shows that two periods of extension with different orientations have occurred in this region. These differing extension directions explain the modern distribution of ranges and valleys and the physiography of the ranges in the entire Inyo-Death Valley region. Middle Miocene extension produced northwest-trending normal faults with offsets of 1 km or more and a relatively simple northwest-trending basin-and-range topography. A period of relative tectonic quiescence during which the uplifted blocks were eroded and sediment began to fill the basins followed this uplift. Extensional faulting in the Pliocene, beginning soon after 4 Ma, was considerably more complicated due to the reorientation of the extension direction. Accommodation of this new extension direction took place either by offset on the older, "inherited" structures, or by the formation of new faults. Extension in the Saline Range was accommodated by a series of closely spaced, northeast-trending normal faults. In the Inyo Mountains, renewed normal offset on the northern segment of the older EIFZ was accommodated by right-lateral offset on the newly-formed HMFZ and westward tilt of the floor of Saline Valley. Rapid down-drop of Saline Valley led to the formation of the jumble, a tectonic breccia resulting from gravity sliding of the bedrock into the deep Saline Valley depression. The southern segment of the EIFZ was not rejuvenated as shown by 14-Ma gravels that cover the fault. Instead, extension at this latitude was accommodated to the east on the west side of the Panamint Range along a west-dipping, low-angle normal fault that resulted in the opening of modern Panamint Valley by northwest transport of the Lee Flat and Inyo Mountain blocks.

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APPENDIX: ANALYTICAL METHODS

K-Ar Analyses

Conventional K-Ar analyses were done in the U.S. Geological Survey laboratories, Menlo Park, California, using standard isotope-dilution techniques similar to those described by Dalrymple and Lanphere (1969). Relatively pure samples of the tephra were crushed and sized to 105-150 microns (100-140 mesh). A biotite separate was prepared by standard heavy liquid, magnetic, electrostatic, and handpicking procedures. The samples were fused by induction heating and a 60°-sector, 15.2-cm-radius, Neir-type mass spectrometer was used for argon analysis. Potassium analyses were performed by a lithium metaborate flux fusion-flame photometer technique, using lithium as an internal standard (Ingamells, 1970).

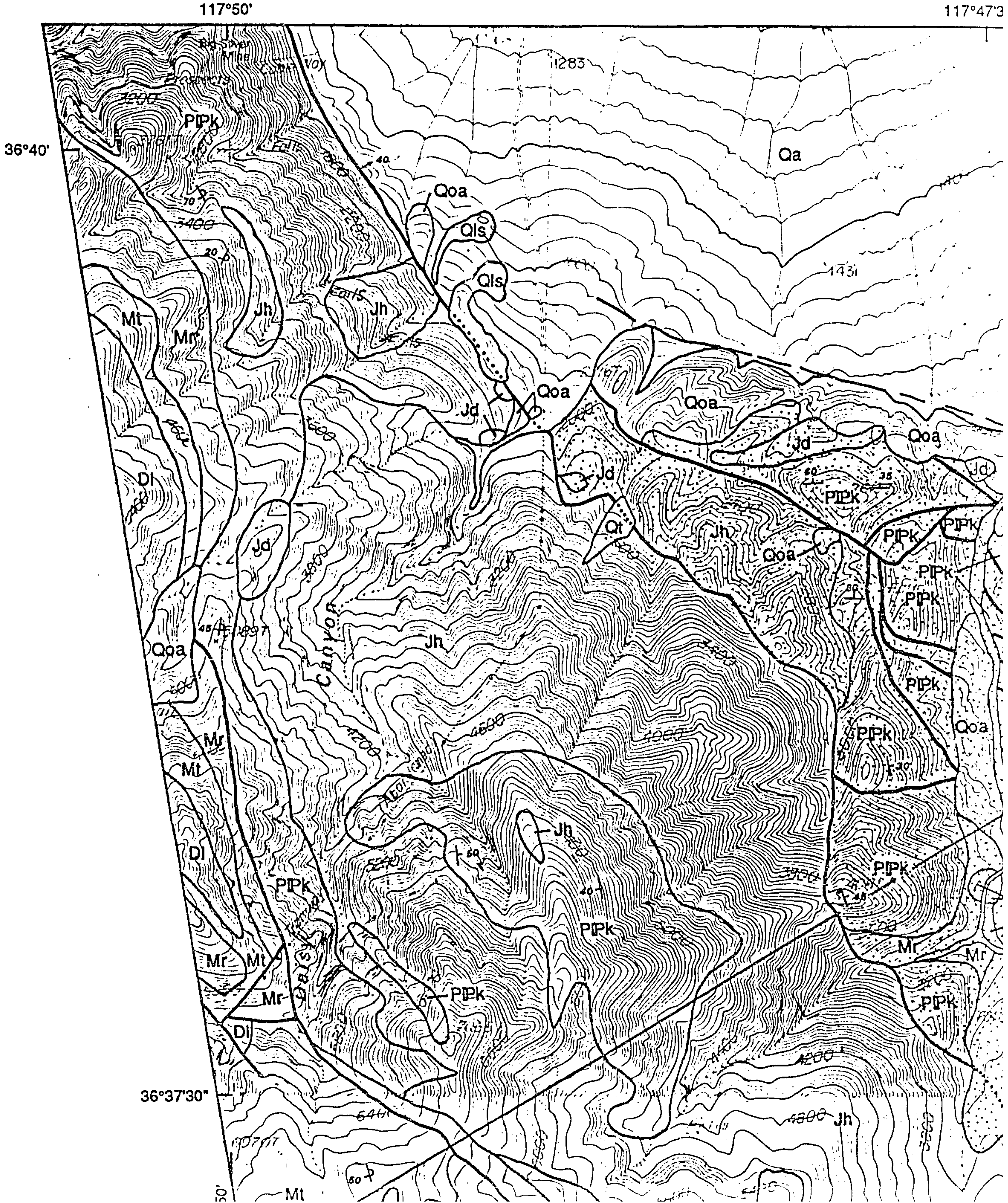
$^{40}Ar/^{39}Ar$ Analyses

Samples used for the $^{40}Ar/^{39}Ar$ analyses were irradiated in the Los Alamos Omega West reactor for 8 hours. The J-value, a measure of the irradiation flux, was calculated by use of standards MMhb-1 hornblende, with an age of 520.4 Ma, GHC 305 biotite, with an age of 103.76 Ma, and P-207 muscovite, with an age of 82.1 Ma. The following potassium and calcium corrections for the Omega West reactor were determined using optical grade CaF_2 and a laboratory potassium glass: $^{40}Ar_K/^{39}Ar_K$ = 0.0024, $^{39}Ar_{Ca}/^{37}Ar_{Ca}$ = 0.000729, and $^{36}Ar_{Ca}/^{37}Ar_{Ca}$ = 0.000275.

The laser-heated $^{40}Ar/^{39}Ar$ extractions and isotopic analyses were conducted at the Berkeley Geochronology Center using a laser-fusion microextraction system and an online, ultrasensitive mass spectrometer. Argon backgrounds for this system are approximately the following: $^{40}Ar = 4.0 \times 10^{-12}$ cc (STP), $^{39}Ar = 1.0 \times 10^{-13}$ cc (STP), $^{37}Ar = 7.6 \times 10^{-14}$ cc (STP), and $^{36}Ar = 3.9 \times 10^{-14}$ cc (STP). Detection limit for this mass spectrometer is on the order of 1.0×10^{-14} cc (STP). Use of laser-light energy for fusion greatly reduces the background argon because there is less nonfocused heating and subsequent outgassing of the ultrahigh vacuum system. This system is designed to analyze samples as small as single grains (less than 0.1 mg) depending on the total argon content as determined by grain size, age, and potassium content. Approximately 3 to 5 grains of biotite were used in each experiment to provide sufficient gas for accurate analysis.

The precision of the calculated age, shown as the ± value, is the estimated analytical uncertainty at one standard deviation. It represents uncertainty in the measurement of the argon isotopes, radiogenic ^{40}Ar yield, and K_20, and is confirmed by replicate analyses of mineral standards. For the analyses, the J-value error is assumed to be 0.5 percent based on replicate analyses of the standards. The decay constants used for K and the abundance ratio of $^{40}Ar/K_{tot}$ are those adopted by the International Union of Geological Sciences Subcommission on Geochronology (Steiger and Jäger, 1977).

The development of new, ultrasensitive mass spectrometers during the last 10 years (York and others, 1981) now permits argon analysis of samples weighing less than 0.1 mg. In contrast, standard K-Ar techniques require sample sizes of at least 250-400 mg for argon analysis and 100 mg for potassium analysis. The ability to measure argon from extremely small samples using these techniques now permits analysis of single grains or small populations of grains to evaluate possible grain to grain variation in age in a given sample. These techniques are now commonly used to detect the presence of older contaminant mineral grains in pyroclastic volcanic rocks that may lead to anomalously old K-Ar ages.



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36°40'
36°37'30"
Big Silver Mine
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Qoa
Qls
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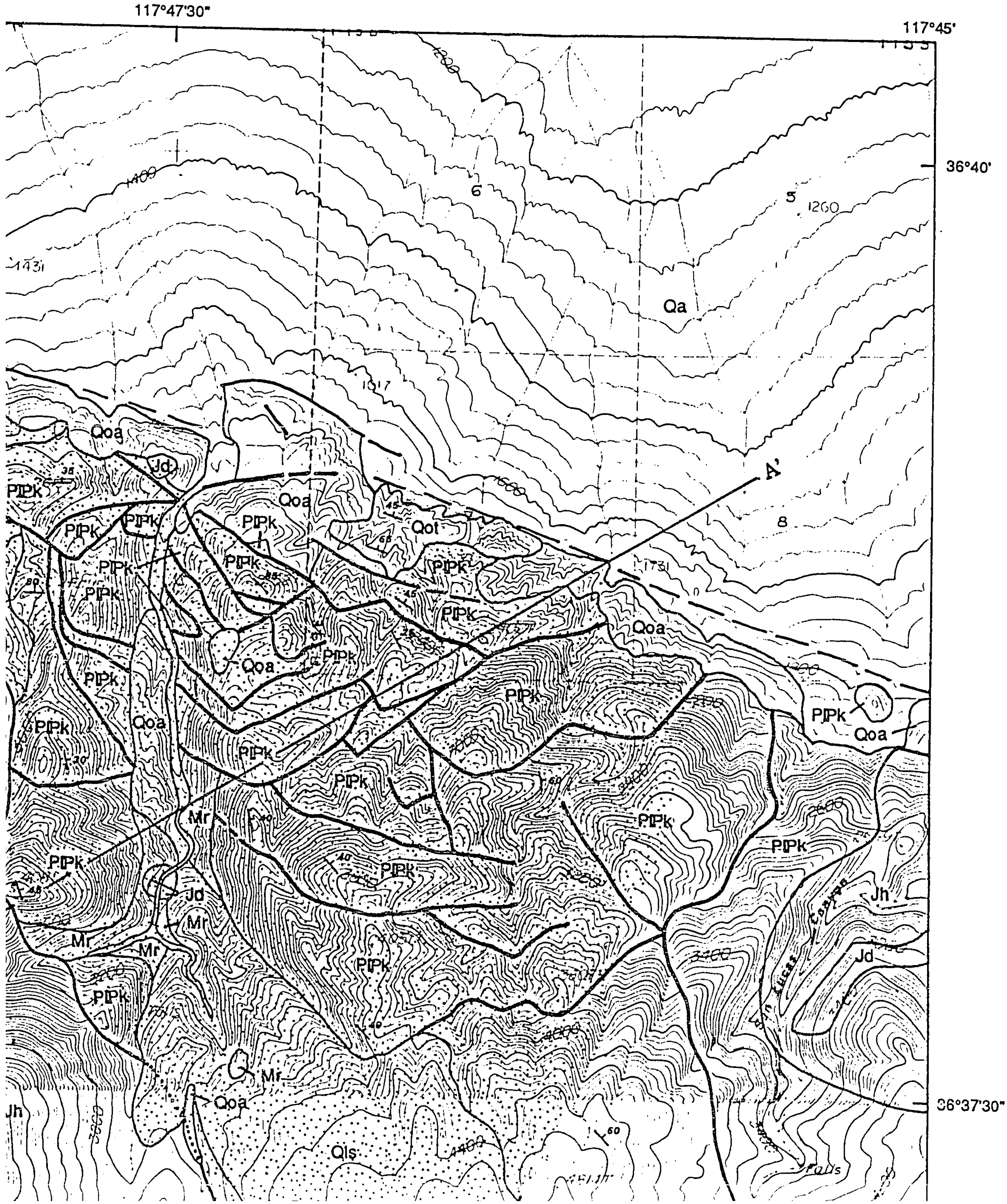

117°47'30"
117°45'
36°40'
36°37'30"
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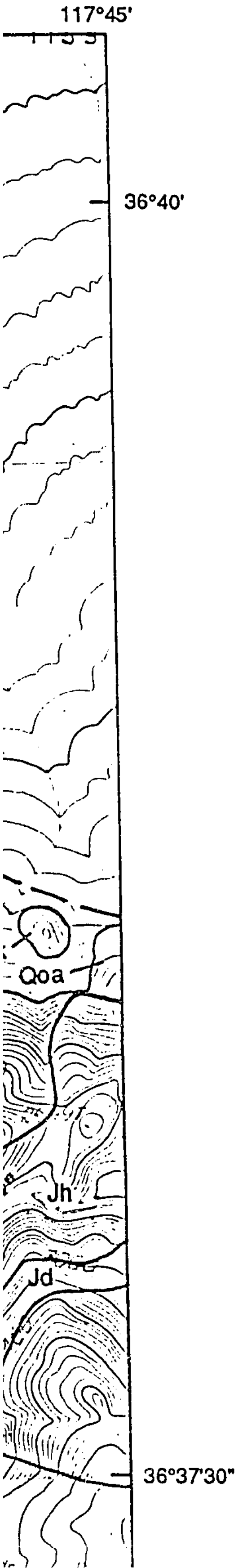

CORRELATION OF MAP UNITS

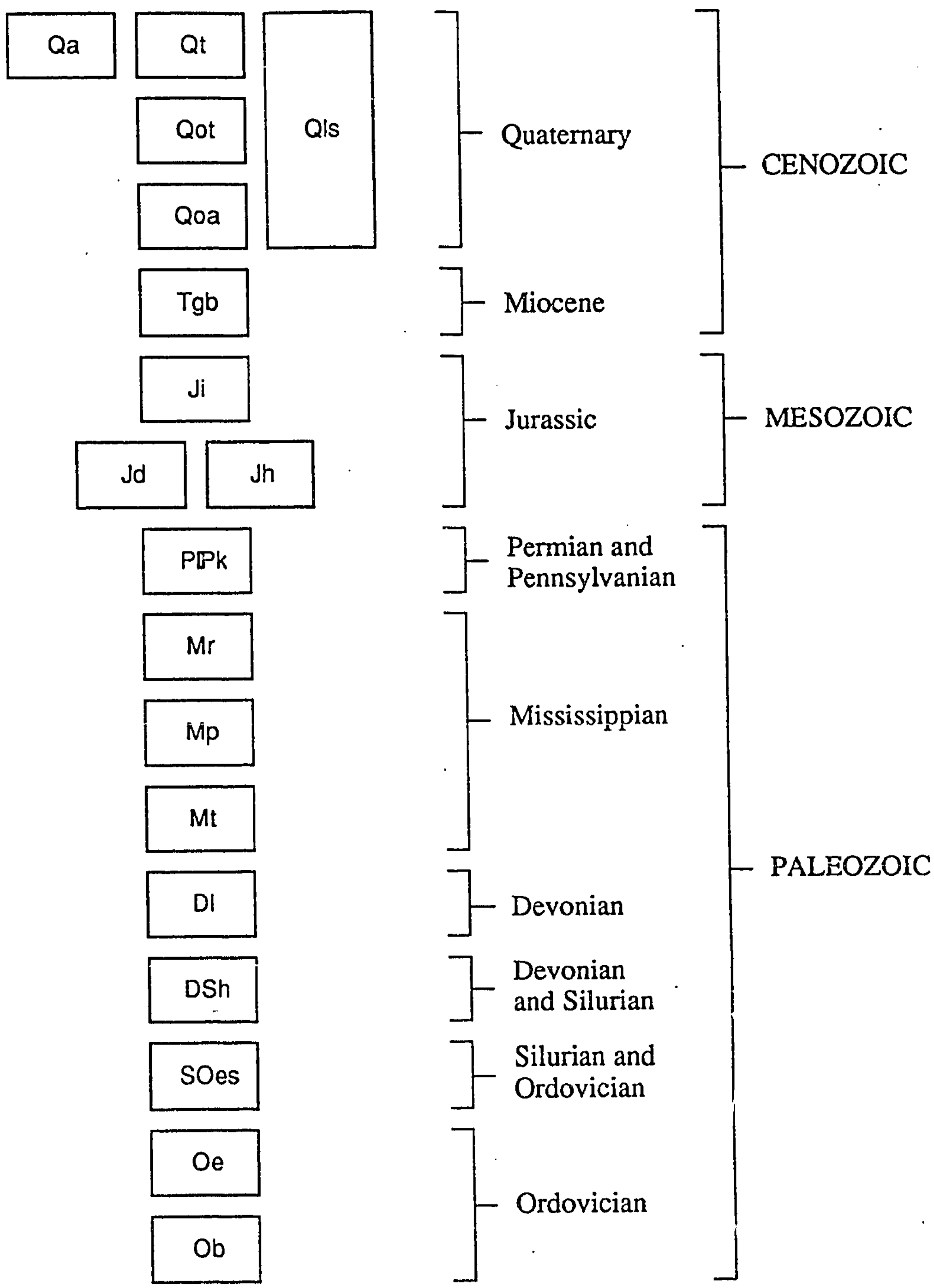

PLATE 1

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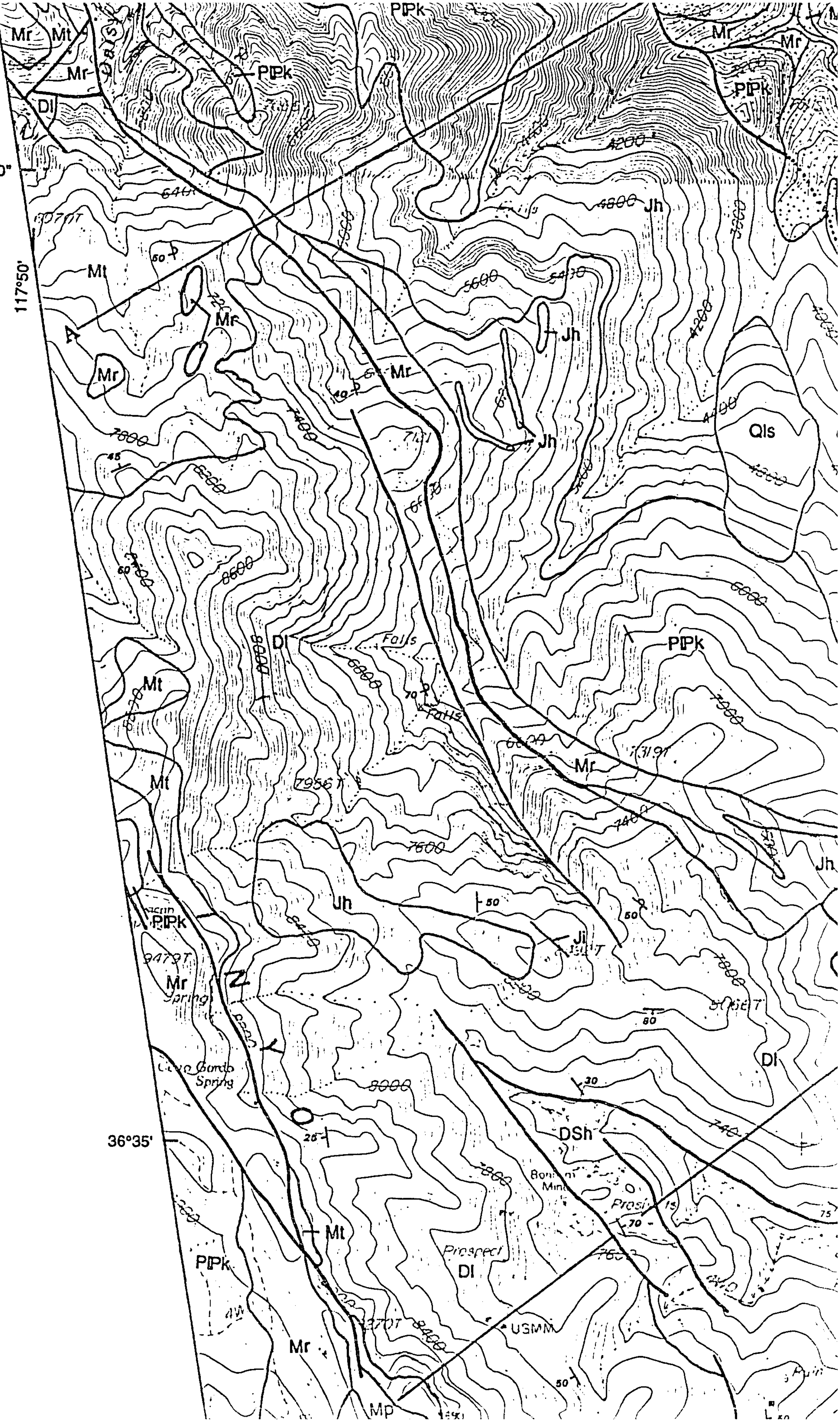
DESCRIPTION OF MAP UNITS

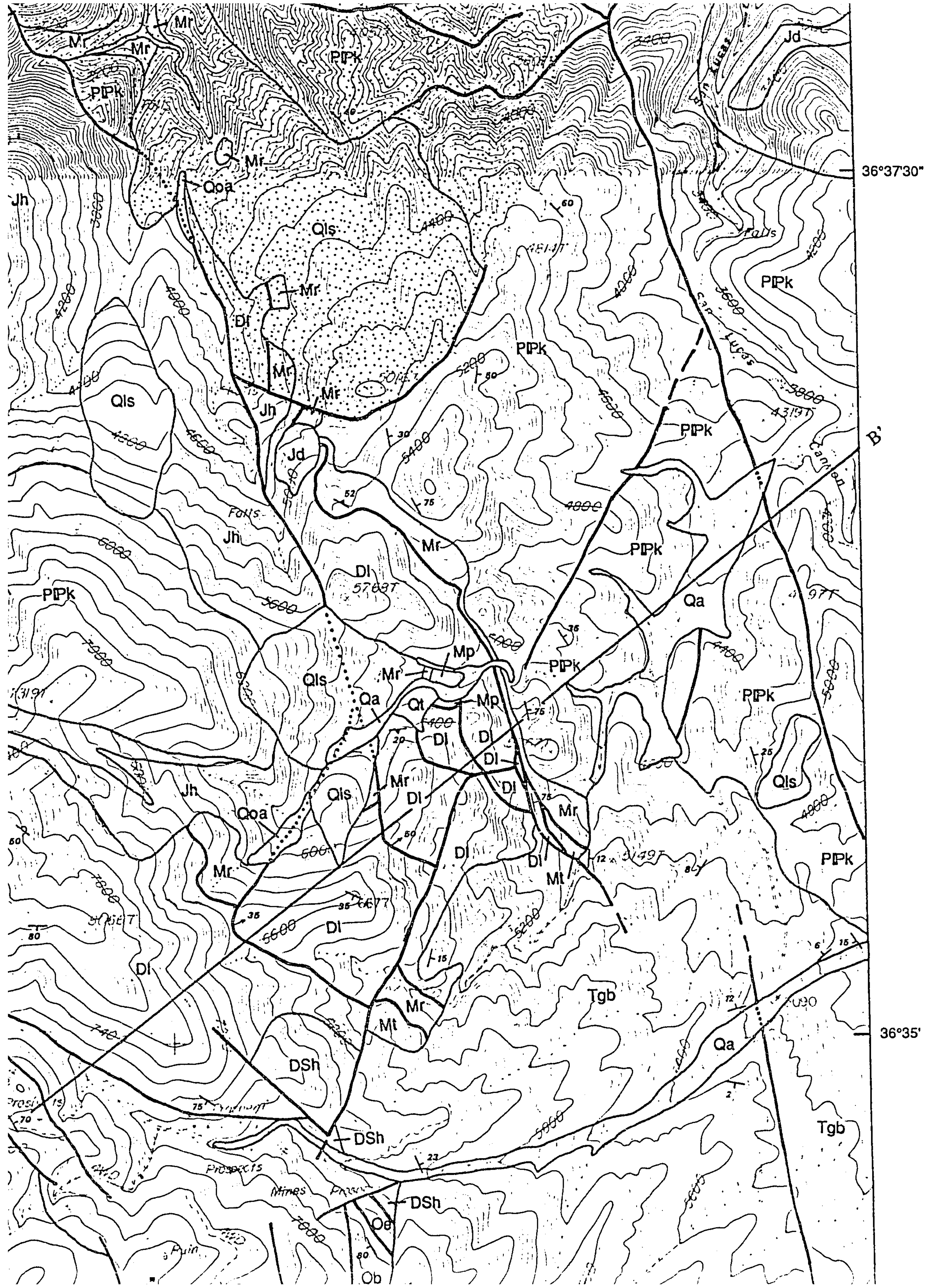
Qa	**Alluvium (Quaternary)**--Unconsolidated sand and gravel
Qt	**Talus (Quaternary)**
Qls	**Landslide Deposits (Quaternary)**
Qot	**Older Talus (Quaternary)**--Cemented talus deposits
Qoa	**Older Gravels (Quaternary)**--Cemented fanglomerate and stream deposits
Tgb	**Gravels of Bonham Canyon (Tertiary)**--Well-bedded to massive pebble, cobble, and boulder conglomerate
Ji	**Independence Dike Swarm (Jurassic)**--Mostly mafic and typically altered dikes ranging in composition from granite to diorite
Jh	**Hunter Mountain Quartz Monzonite (Jurassic)**--Medium-grained, seriate quartz monzonite, gradational to medium- to fine-grained, dark-colored rocks ranging in composition from quartz monzonite to diorite
Jd	**Diorite (Jurassic)**--Fine- to medium-grained diorite
PIPk	**Keeler Canyon Formation (Permian and Pennsylvanian)**--Thick beds of bioclastic, blue-gray limestone alternating with thin-bedded, limy shale
Mr	**Rest Spring Shale (Mississippian)**--Dark-gray to black, argillaceous shale with minor fine-grained sandstone and siltstone
Mp	**Perdido Formation (Mississippian)**--Tan to medium- and dark-gray limestone, siliceous limestone, siltstone, chert, and quartzite
Mt	**Tin Mountain Limestone (Mississippian)**--Dark-blue, thin- to thick-bedded limestone
Dl	**Lost Burro Formation (Devonian)**--Massive white to bluish-gray limestone and marble
DSh	**Hidden Valley Dolomite (Devonian and Silurian)**--Massive white to tan, dolomitic marble, well-bedded, cherty dolomite, and massive quartzite
SOes	**Ely Springs Dolomite (Silurian and Ordovician)**--Dark blue-gray, medium-grained dolomite with abundant chert blebs and stringers
Oe	**Eureka Quartzite (Ordovician)**--Thick-bedded to massive, medium-grained, white, vitreous orthoquartzite
Ob	**Badger Flat Limestone (Ordovician)**--Blue-gray, sandy and silty

36°37'30"

117°50'

36°35'





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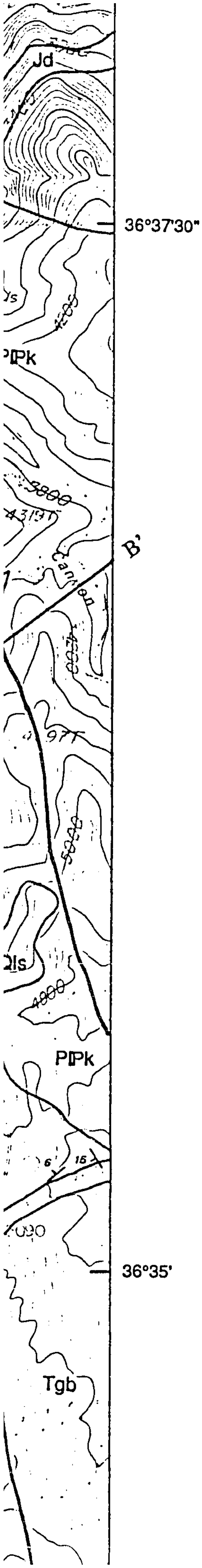

Ordovician

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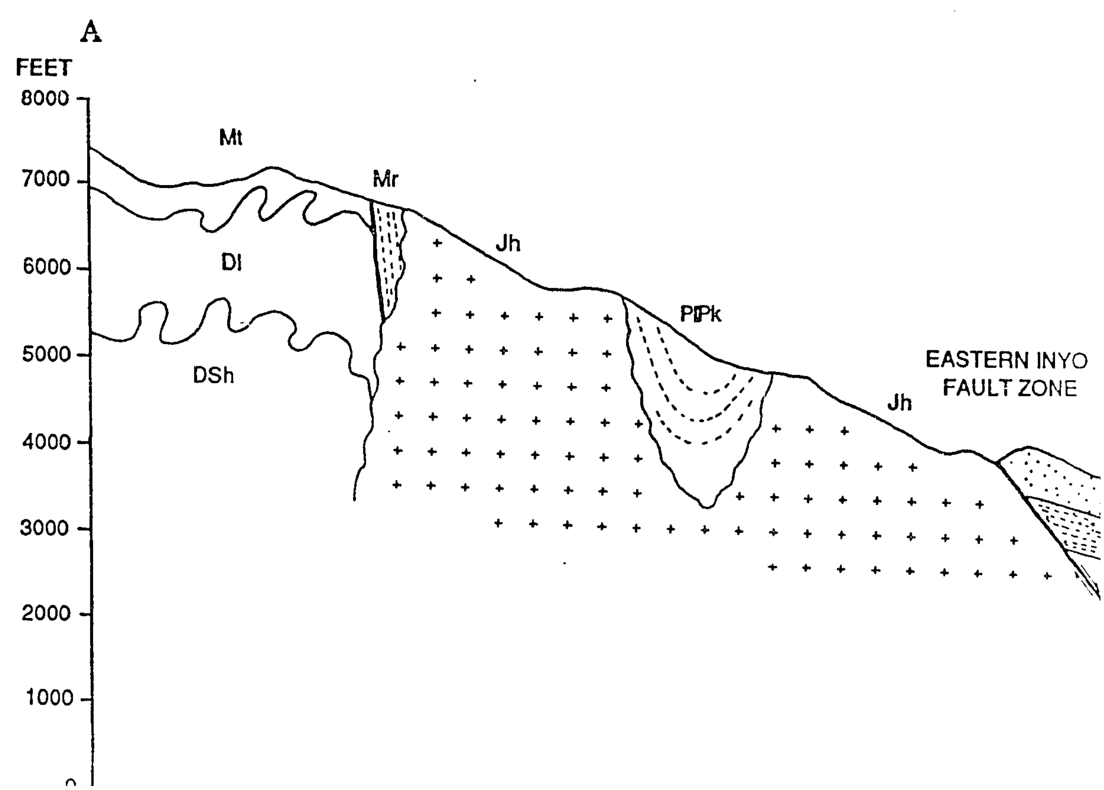
Ordovician



bedded limestone

Dl	**Lost Burro Formation (Devonian)**--Massive white to bluish-gray limestone and marble
DSh	**Hidden Valley Dolomite (Devonian and Silurian)**--Massive white to tan, dolomitic marble, well-bedded, cherty dolomite, and massive quartzite
SOes	**Ely Springs Dolomite (Silurian and Ordovician)**--Dark blue-gray, medium-grained dolomite with abundant chert blebs and stringers
Oe	**Eureka Quartzite (Ordovician)**--Thick-bedded to massive, medium-grained, white, vitreous orthoquartzite
Ob	**Badger Flat Limestone (Ordovician)**--Blue-gray, sandy and silty limestone and siltstone with abundant chert in the lower part

Contact

28

Fault--Showing dip; dashed where approximately located, dotted where concealed

Strike and dip of beds

10 Inclined

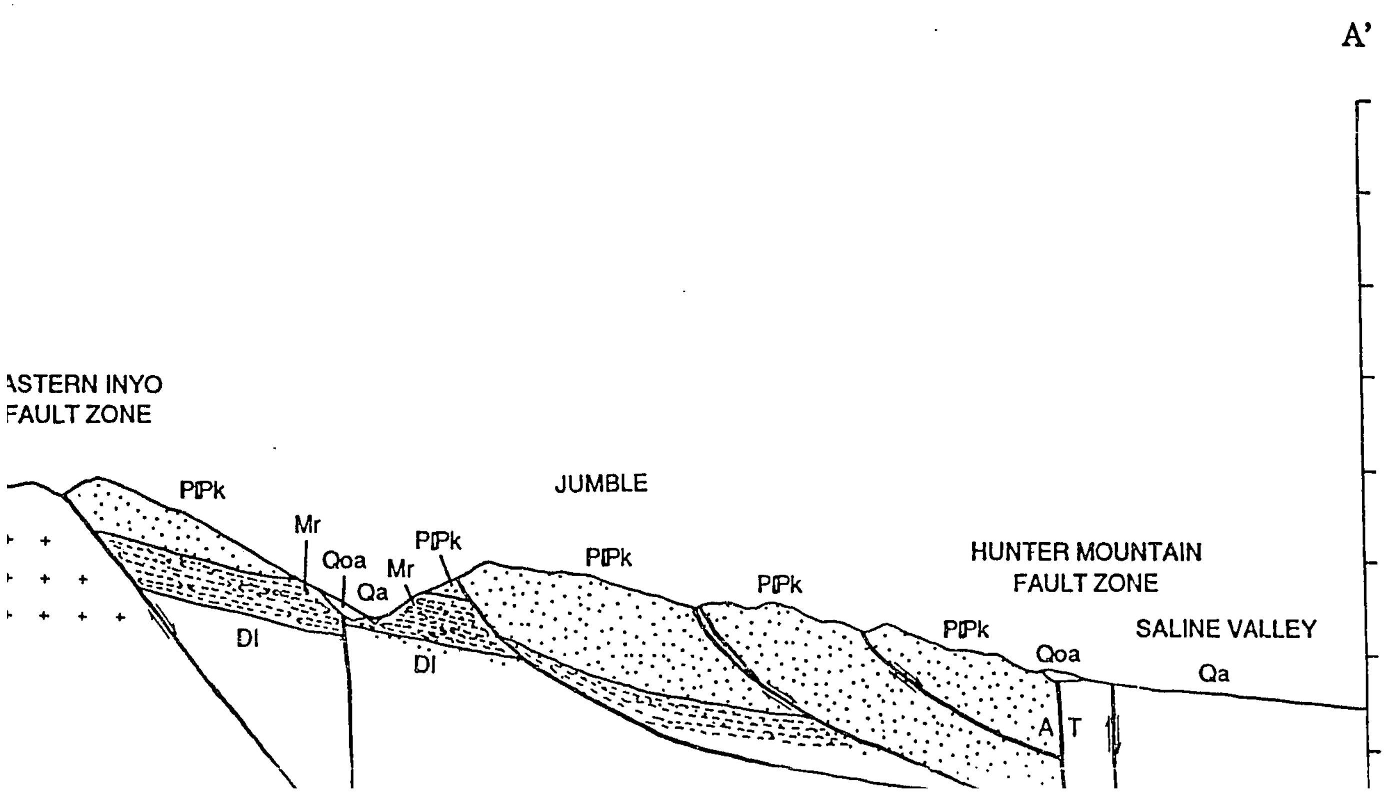
Vertical

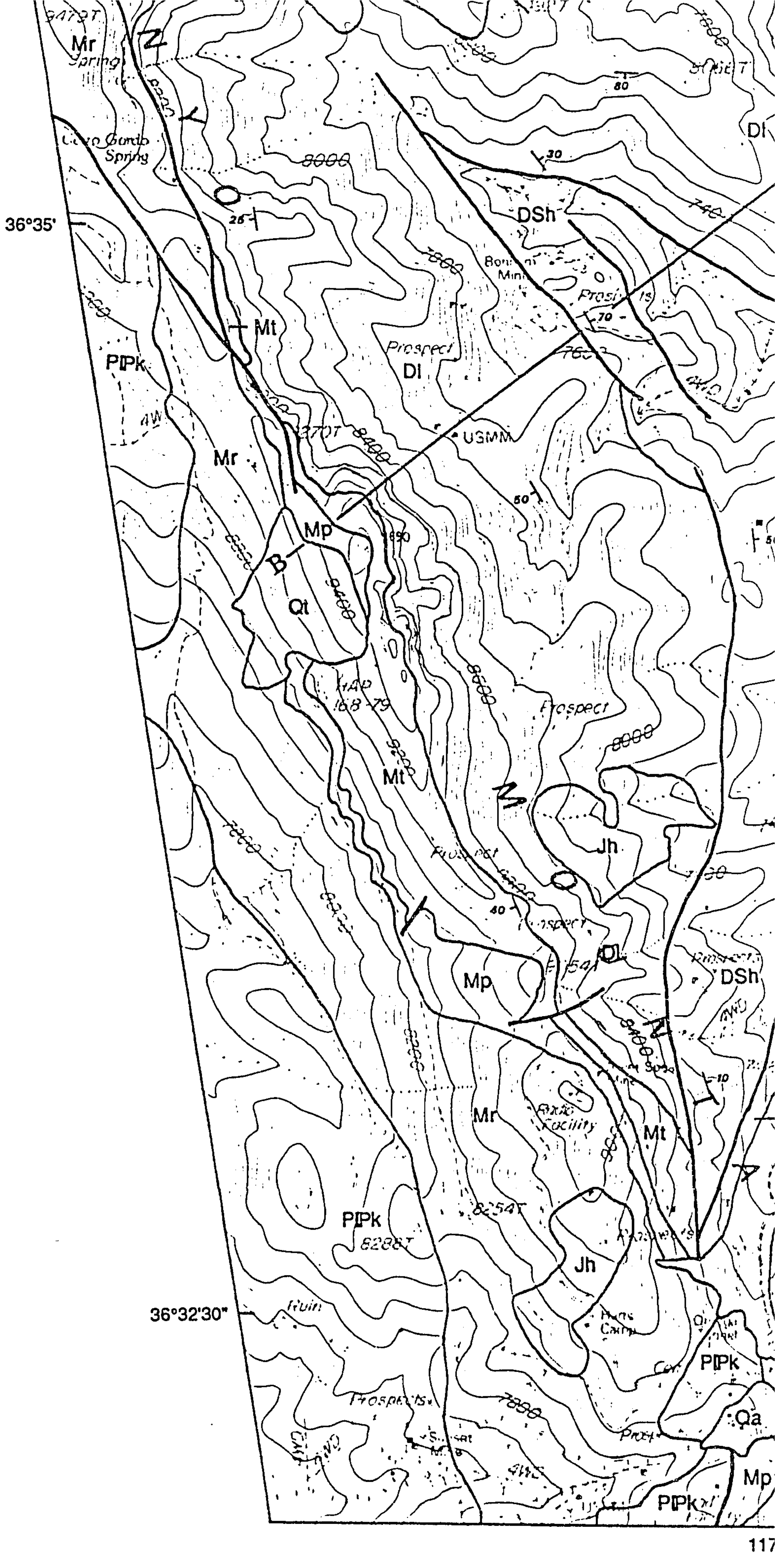
60 Overturned

Strike and dip of foliation

17 Inclined

Jumble





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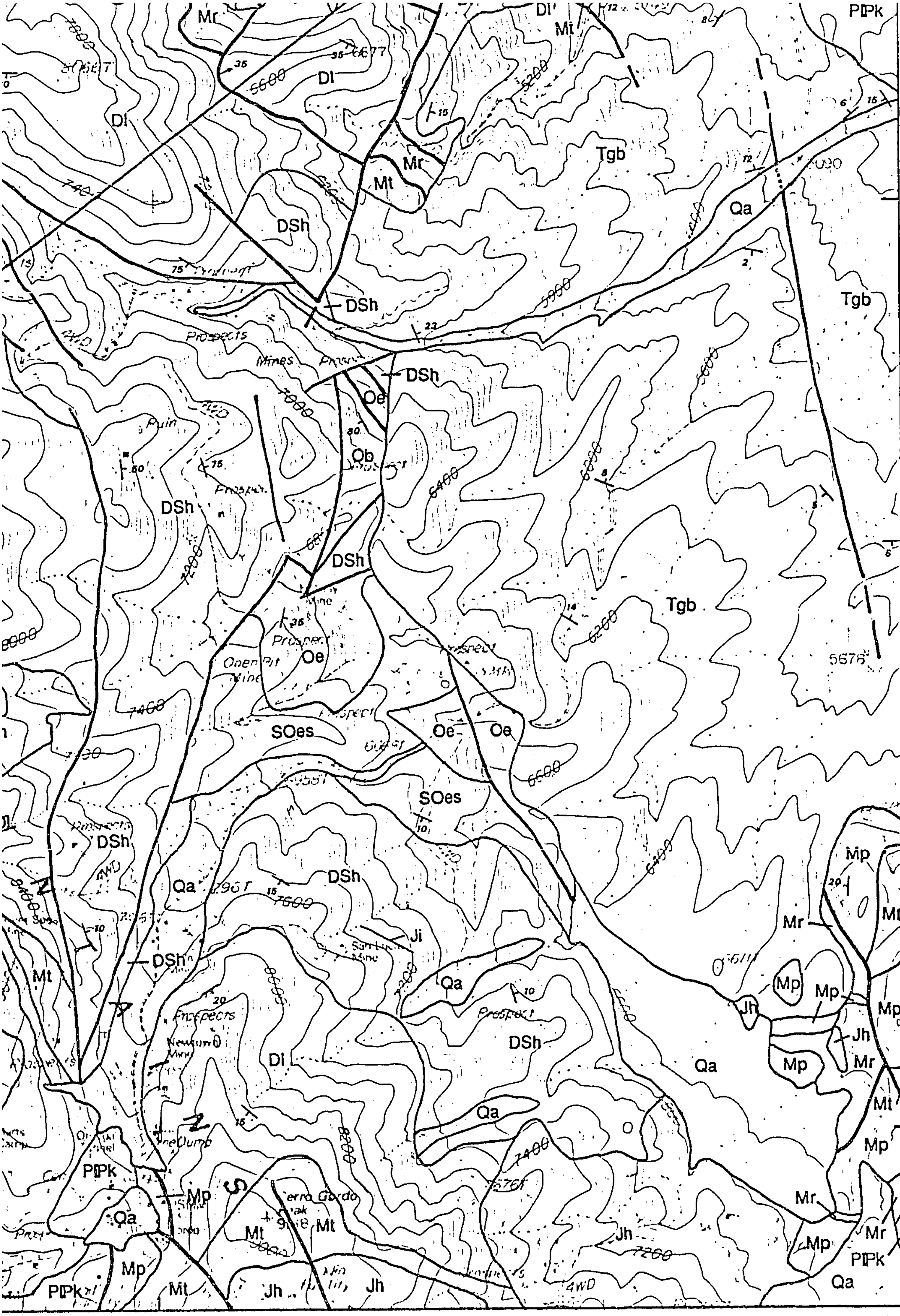
Cerro Gordo Peak, Craig Canyon 1987



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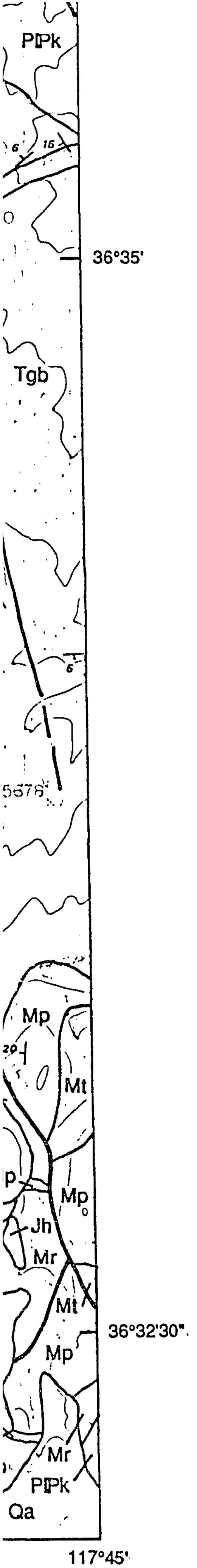

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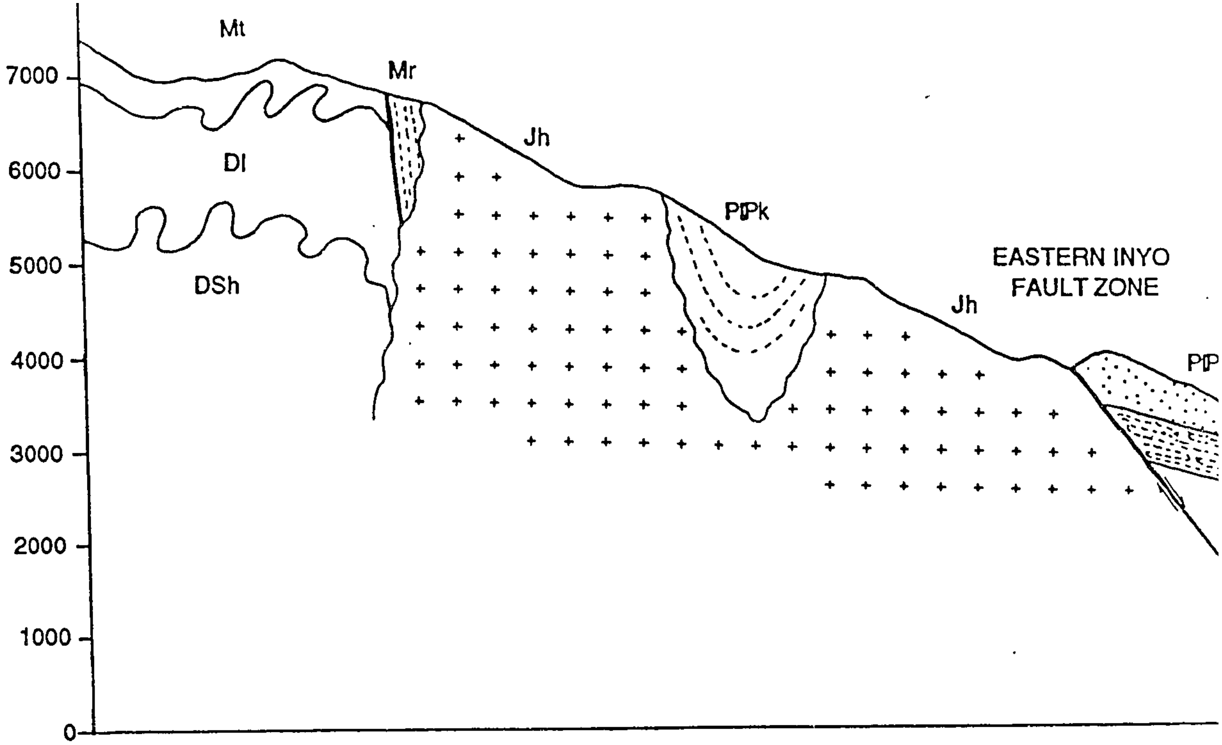
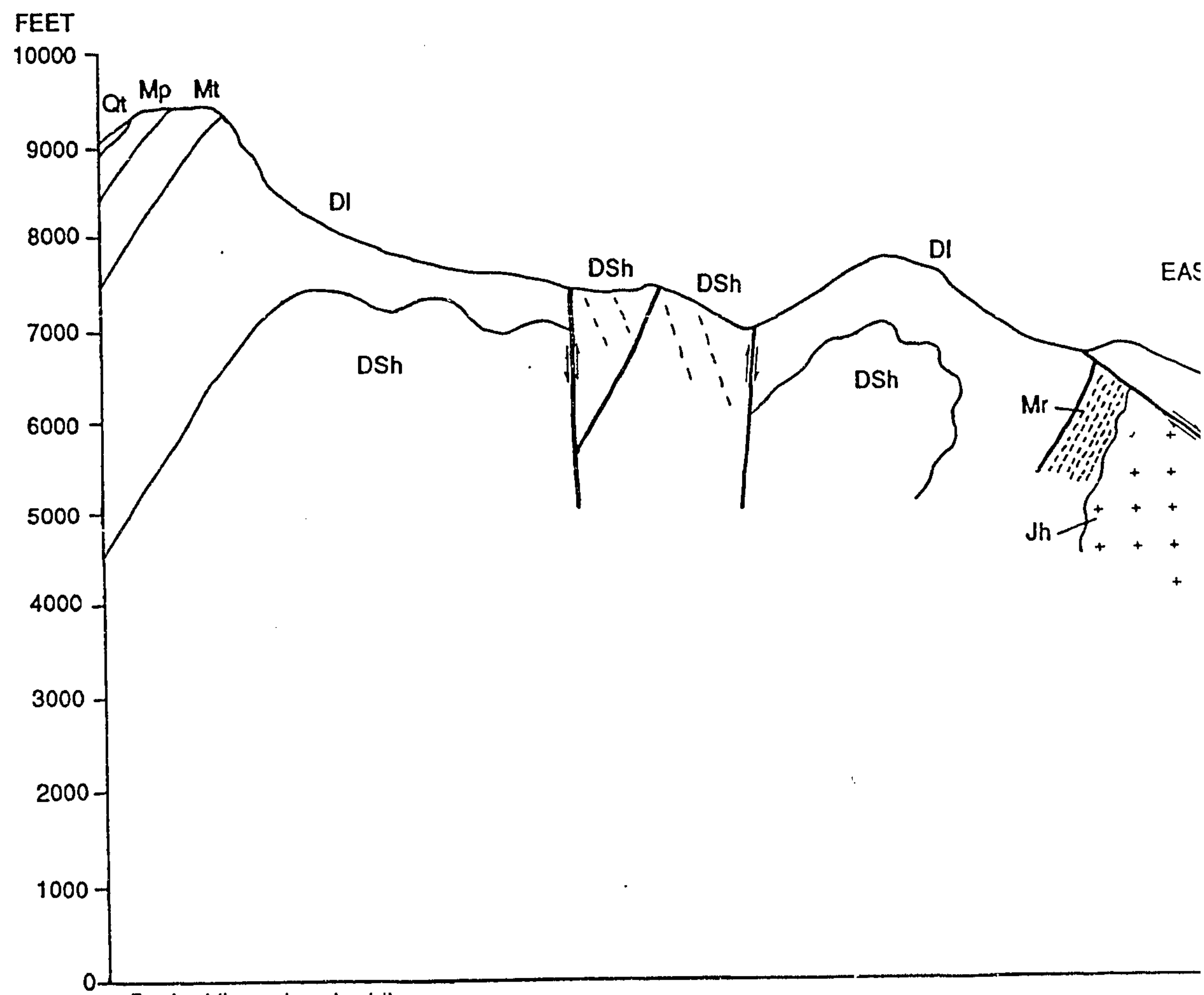


T, toward viewer A, away from viewer

Dashed lines show bedding

B



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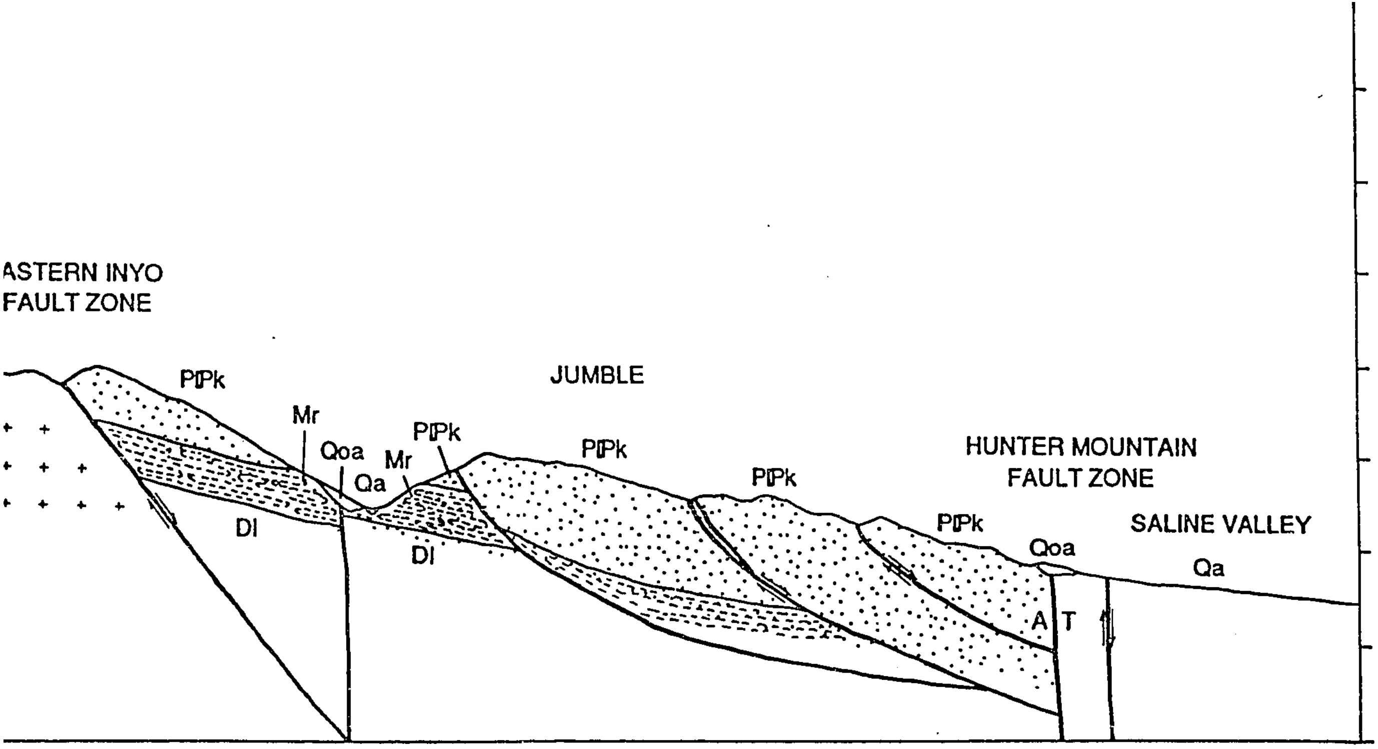
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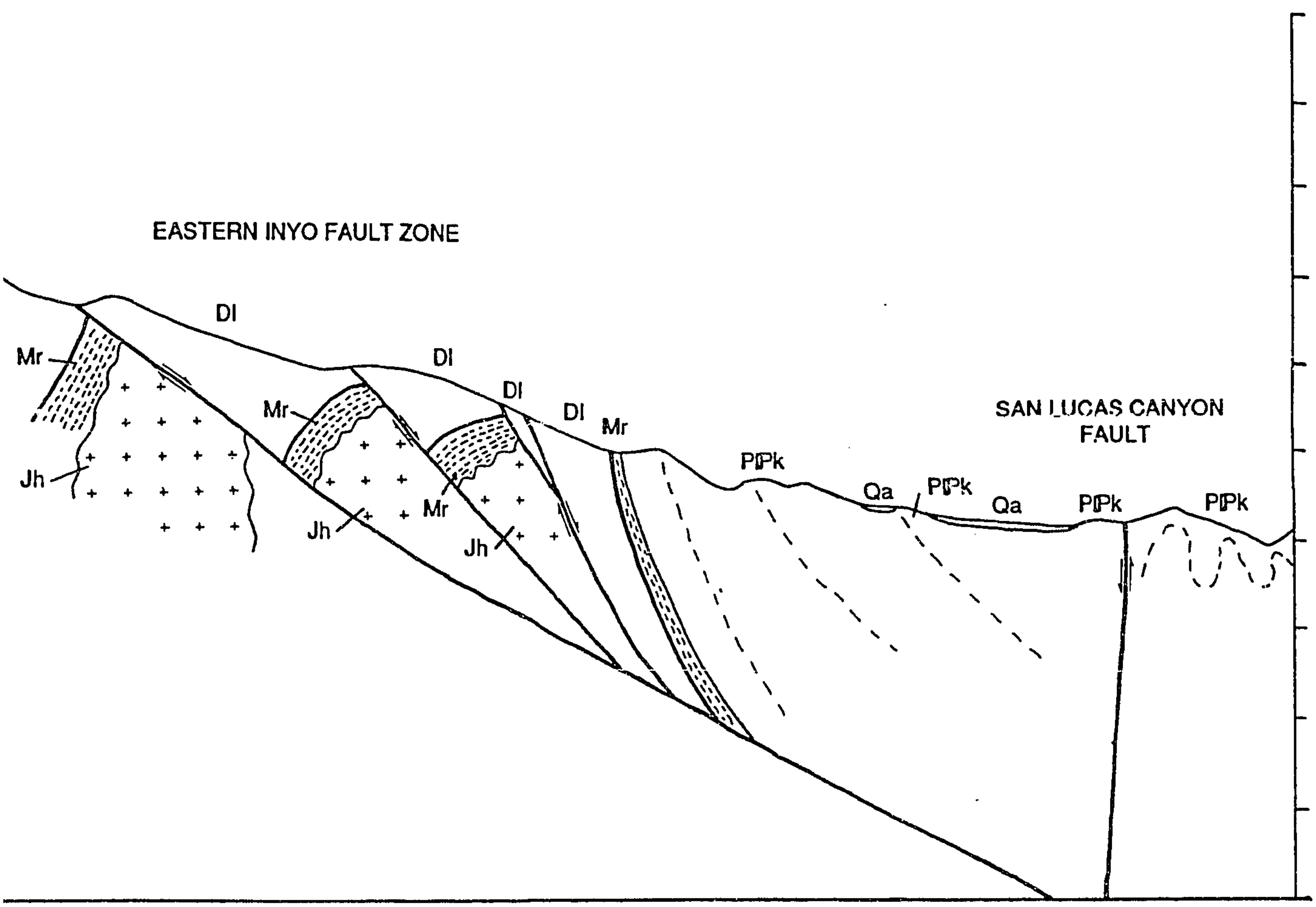
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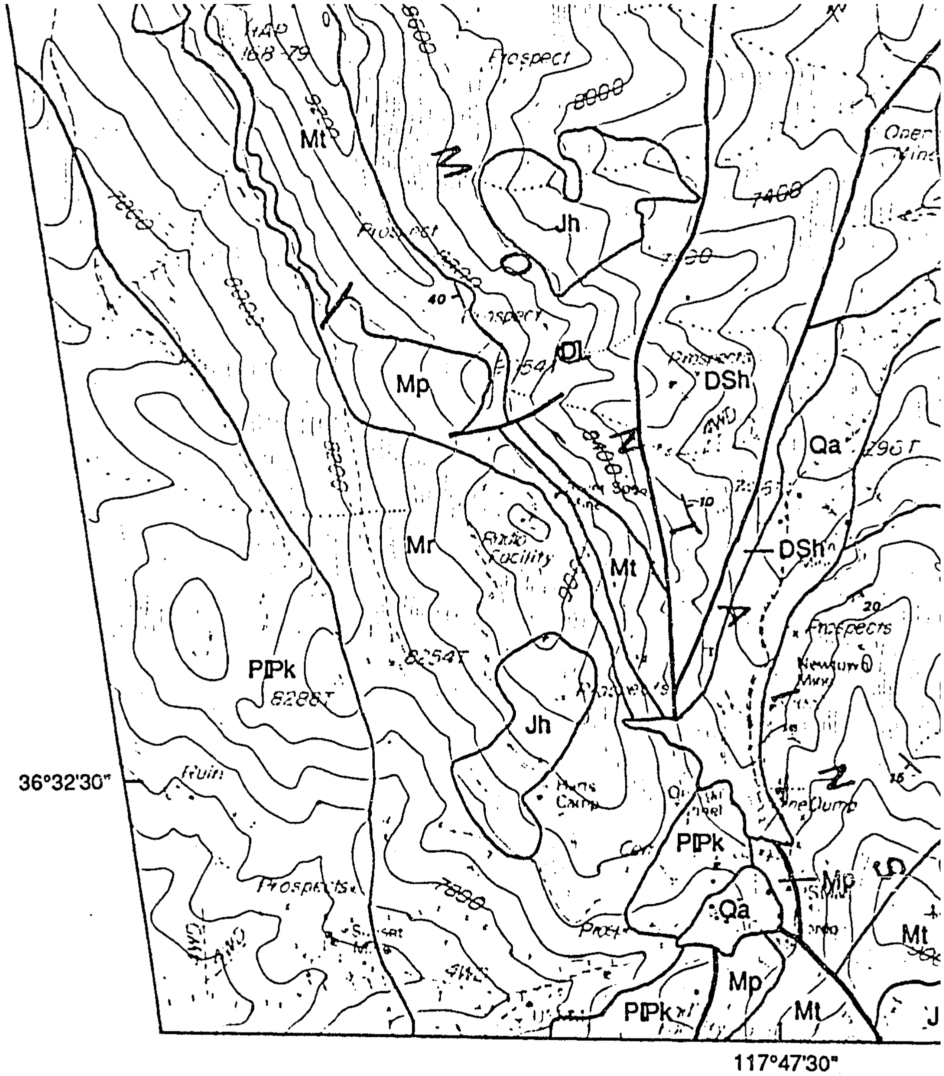

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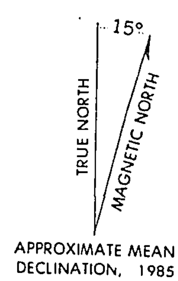
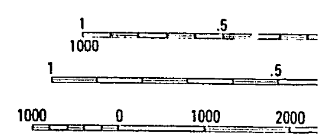
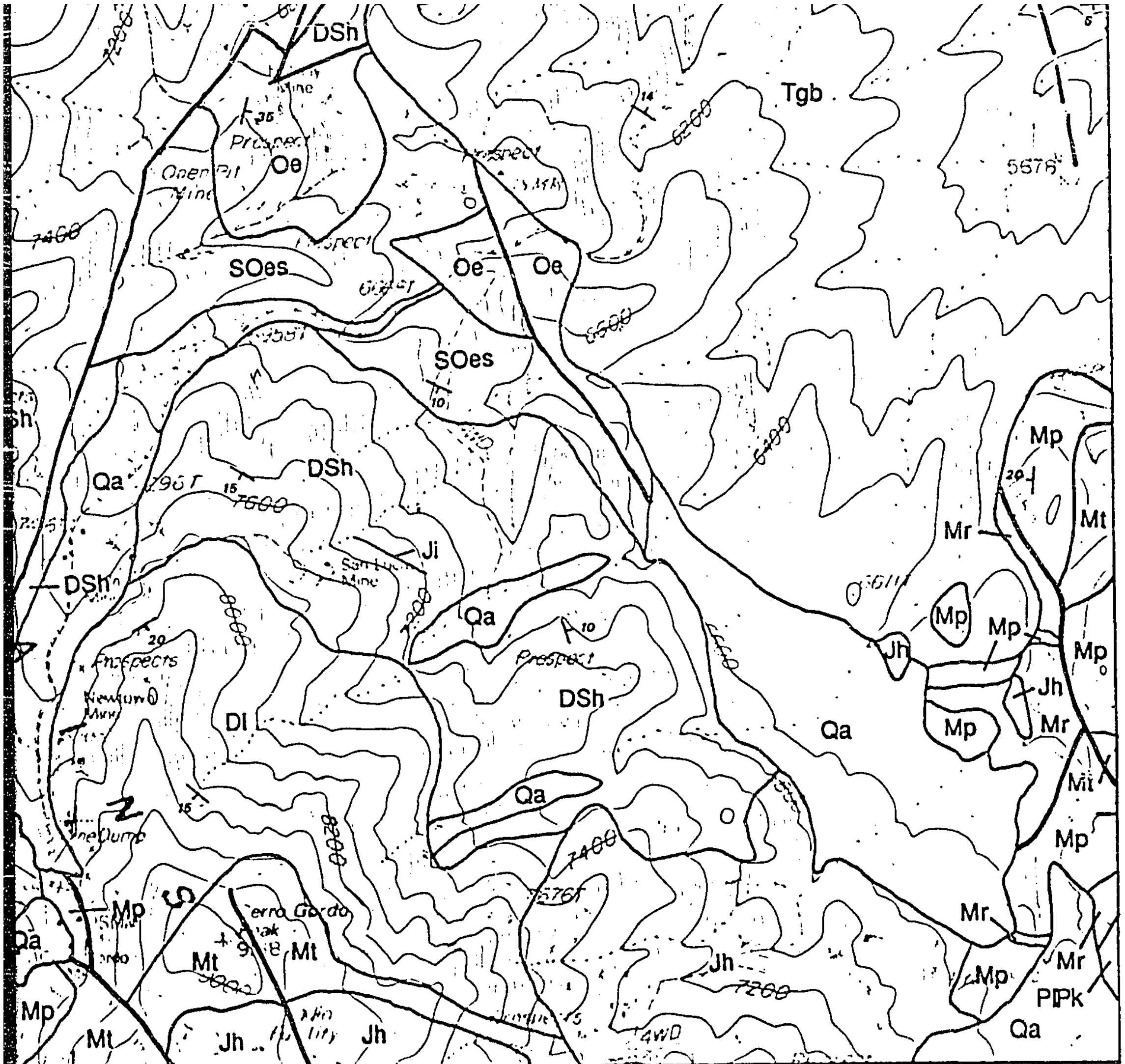
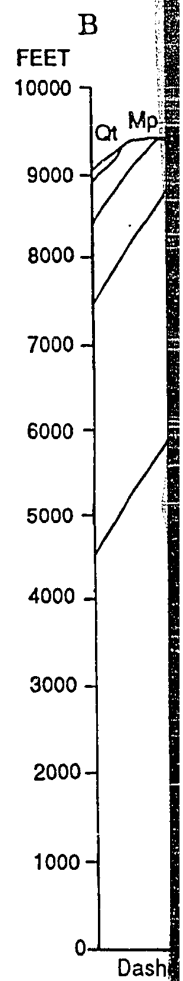
Cerro Gordo Peak, Craig Canyon 1987









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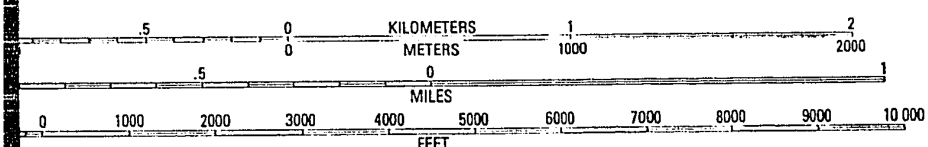
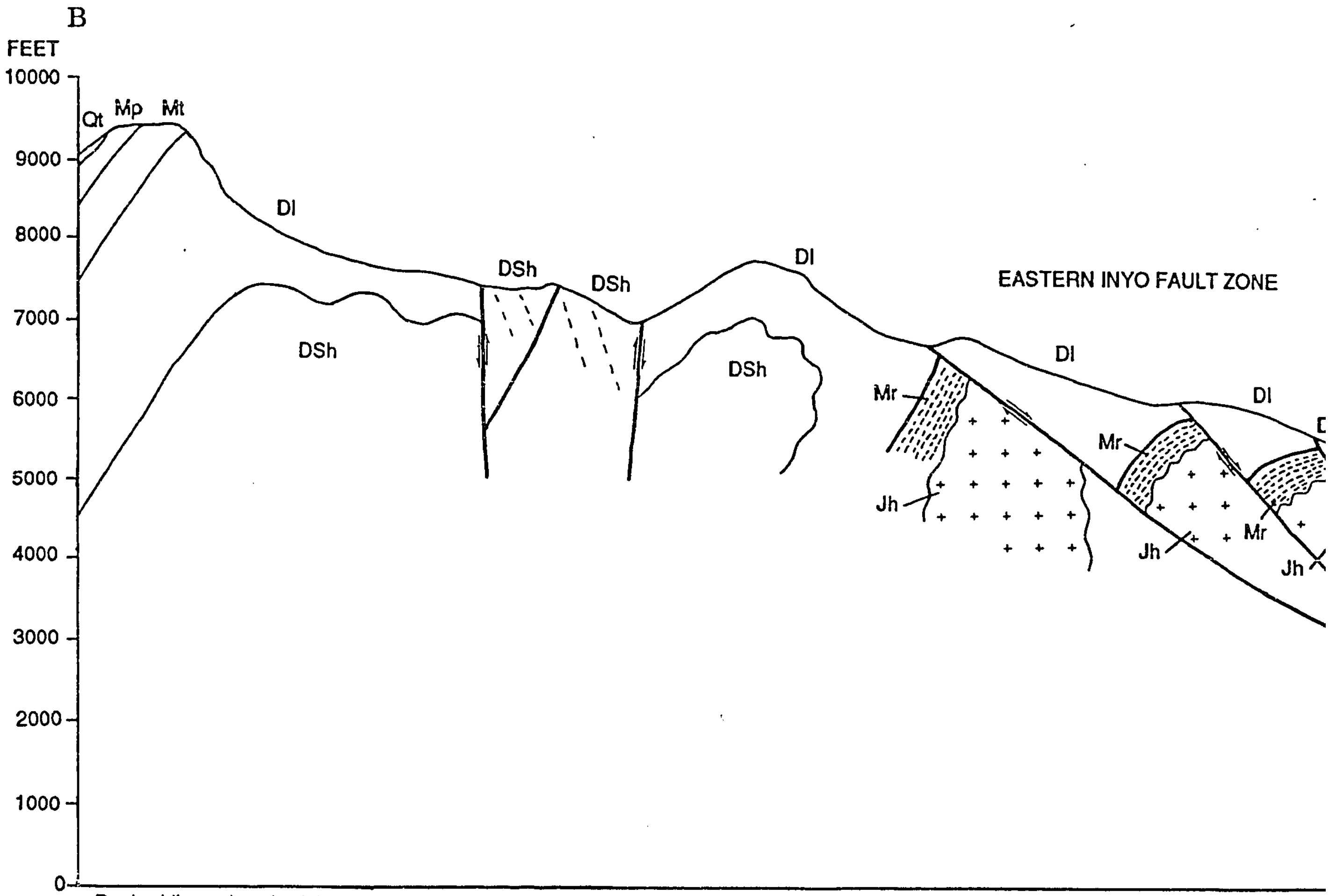


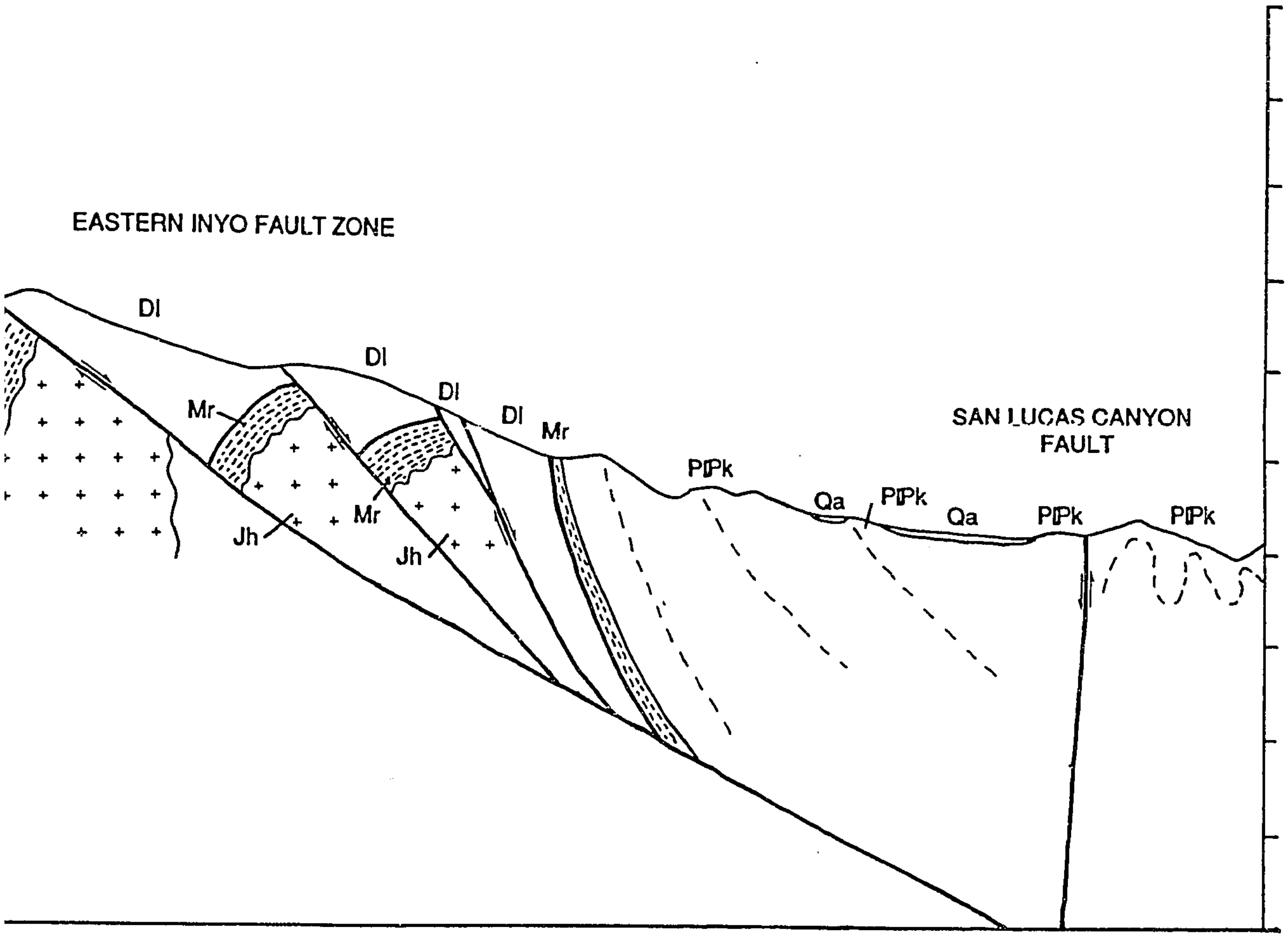
PLATE 1. GEOLOGIC MAP OF A PORTIC
MOUNTAINS, INYO COU



Dashed lines show bedding

\ PORTION OF THE SOUTHEASTERN INYO
IYO COUNTY, CALIFORNIA

B'


EASTERN INYO FAULT ZONE
Dl
Dl
Dl
Dl
Mr
Mr
Mr
Jh
Jh
PIPk
Qa
PIPk
Qa
PIPk
PIPk
SAN LUCAS CANYON
FAULT

O